AERONAUTICAL METEOROLOGY

by

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FOREWORD

Air transportation is today successful; not entirely for the reason that airplanes are structurally sound, or that they are fast, or that they are pleasing to the eye. The success of this form of transportation, in which we are all interested, is founded upon public acceptance, and public acceptance is, in turn, founded upon public confidence.

People have confidence in knowledge, and the meteorologist has made a definite and tangible contribution to the sum of knowledge that we have concerning the safe operation of aircraft.

No factor in safe flight is more important than knowledge of existing weather conditions and knowledge of probable trends. Safety of operation of aircraft under existing methods would be impossible without the aid of the meteorologist.

We know more about meteorology than we did some years ago. That knowledge has permitted us to broaden the usefulness of the service which is performed by aircraft. We acknowledge and appreciate the contribution which the science of meteorology has made to the progress and success of air transportation. We look forward to further horizons of service, firm in the belief that the aid which has permitted us to attain the goal of today will permit us to be safe in striving for one more distant.

George F. Taylor has been a pioneer in the application of the principles of meteorology to the practical problems of air transportation. We welcome his book on this subject and we are hopeful that it will enjoy the success which it deserves.

C. R. Smith
Colonel, Air Service Ferrying Command
Formerly President, American Airlines, Inc.

ence. The student who first attempts the construction of a weather chart according to modern methods, however, often finds it difficult to get a satisfactory start, since little or no material has been published on this very practical phase of meteorology.

The author has attempted throughout to present fairly, all sides of moot questions. When this is not done, it is because of lack of space, and not through any desire to present a "favorite" theory. It is believed that the student should realize the insecurity of many of the present theories of meteorology, for in this way he can develop a critical attitude toward well established principles. Only in this manner is progress possible. Nothing leads to stagnation of scientific endeavor as completely as passive acceptance of old ideas.

A famous educator once remarked, "It is fortunate that a college course lasts only four years, for that is the life of most of our scientific theories." This is as true for meteorology as for other sciences. Many older theories of circulation, storm structure, and atmospheric composition are being discarded, and new ones are being advanced to take their place. At the present time, attention is being shifted to the air mass as the basis of weather analysis and forecasting. It is being realized, in fact, that any discussion of the weather based on the pressure field alone, neglects the actual *causes* of the observed phenomena, and deals only with the *effects*. The mechanics of the pressure field also have been investigated intensively recently, and have proved to be of great value in practical forecasting.

It is to be hoped sincerely that pure research on the dynamics, kinematics, and thermodynamics of the atmosphere will continue, for every important recent advance in meteorology has been intimately related to such investigations. Many problems facing the meteorologist today appear to be practically incapable of solution. If the tenuous beginnings of other sciences are examined in the light of their present day eminence, however, the meteorologist need not feel discouraged as long as he keeps before him their examples of continuous research. It is a far cry from the first feeble sparks of Hertz to the modern radio set, or from the ancient alchemist to the scores of industries founded on chemistry, yet pure research in physics and chemistry has made possible these outstanding achievements.

The meteorologist, at the threshold of rapid development in his science must ponder the lessons of other fields on knowledge, and never be content with the scope of his learning.

> A little learning is a dangerous thing;
> Drink deep, or taste not the Pierian spring:
> There shallow draughts intoxicate the brain,
> And drinking largely sobers us again.
>
> Alexander Pope

PREFACE TO REVISED AND ENLARGED EDITION

Advances in the meteorological field have necessitated a revised and enlarged edition of this book after three years. The elaboration and wide acceptance of what may be called the Bergeron-Findeisen precipitation theory has been an important recent development, and it has been felt desirable to incorporate a new chapter describing the general subject of Condensation and Precipitation.

The Appendix has been enlarged and rearranged slightly to make it more useful. The references at the end of most of the chapters have been brought up to date. Additional material on the practical use of the kinematical formulae has been added to chapter 19. The adoption of the International Numeral Weather Code by the United States Weather Bureau has necessitated a number of minor changes in the text.

Appreciation is expressed for the many criticisms and suggestions received from meteorologists since the appearance of the first edition, and it is hoped that users of the book will continue to assist the author in this manner in removing errors, and in keeping the text material abreast of new developments.

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THE ATMOSPHERE

INTRODUCTION

The atmosphere is the laboratory in which the meteorologist carries on his experiments. This vast laboratory is unique among scientific laboratories in that the experimenter has practically no control over the reactions that occur within it. While the chemist, the physicist, the biologist, and even the geologist, may control to a greater or less extent the conditions of the problem that he is investigating, the meteorologist can do little more than to observe and ponder. His experimental problem is constantly changing. Rarely, if ever, does it repeat itself exactly. It will be the purpose of the present chapter to point out some of the more interesting features of this ocean of air, on whose floor, and in whose lowermost depths, man walks and flies.

STRUCTURE OF THE ATMOSPHERE

The atmosphere includes all of the gas surrounding the earth, extending from the earth's surface to *outer space.* The point at which *outer space* begins is, of course, somewhat indefinite. Practically, it may well be said to begin where the concentration of molecules becomes vanishingly small. At the surface there are about 27×10^{18} molecules per cc, at 30 miles this is reduced to 3.5×10^{16}, at 80 miles to 3×10^{12}, at 150 miles to 3×10^{6}. Finally, at 500 miles there are only 4 molecules per cubic mile. The total thickness of the atmosphere may thus be taken as not over 500 miles.

The Troposphere—The lower few miles, where most of the atmosphere's mass is concentrated, is called the *troposphere.* It is characterized by considerable motion both horizontal and vertical. Within the troposphere, convection causes rather thorough mixing and tends to establish a fairly rapid decrease in temperature aloft. Water vapor and heat from the surface are also carried aloft to a

considerable elevation. Practically all weather phenomena occur in the troposphere.

The Stratosphere—The convective activity of the troposphere decreases upward and finally ceases entirely at the *tropopause.* Above this, vertical currents are lacking and the temperature no longer decreases. The region above the tropopause is called the *stratosphere.* As its name indicates, the air is stratified here, with the only motions of a horizontal nature. The elevation of the tropopause (or base of the stratosphere) varies from 26,000 to 56,000 feet. It is higher over the equator than over the poles, and somewhat higher during summer than winter. It is also higher over anticyclones than over cyclones.

Within the lower stratosphere, the temperatures are relatively constant. In the higher latitudes, especially, there is little temperature change within a considerable distance above the tropopause. In the lower latitudes there is a marked temperature *increase* above the tropopause. Since the temperatures at the *base* of the stratosphere are *lower* over the equator, this increase aloft causes remarkable uniformity of temperature at all latitudes at elevations near and above 70,000 feet.

Figure 1, which is adapted from Bjerknes, indicates some of the properties of the stratosphere mentioned above. It is interesting to note the cold base of the stratosphere over the equator, the latitudinal variations in the height of the tropopause, the increase in stratosphere temperature above the tropopause over the equator, and the uniformity in temperature within the higher levels of the stratosphere.

The upper winds, which in general increase aloft in the troposphere, reach their maximum velocity near the tropopause, then decrease upward into the stratosphere. Figure 2 indicates average annual wind velocities at various heights near latitude 40°. The number of soundings above 30,000 feet is relatively small but sufficient are available to indicate clearly the general trend in the lower stratosphere.

Average pressures at various heights above the surface are shown in figure 3. The limits of the elevation of the tropopause are indicated here also.

Recent investigations indicate that the upper stratosphere (from 18 miles to 60 miles above the surface) is characterized by increasing temperatures. Experiments on the propagation of sound waves

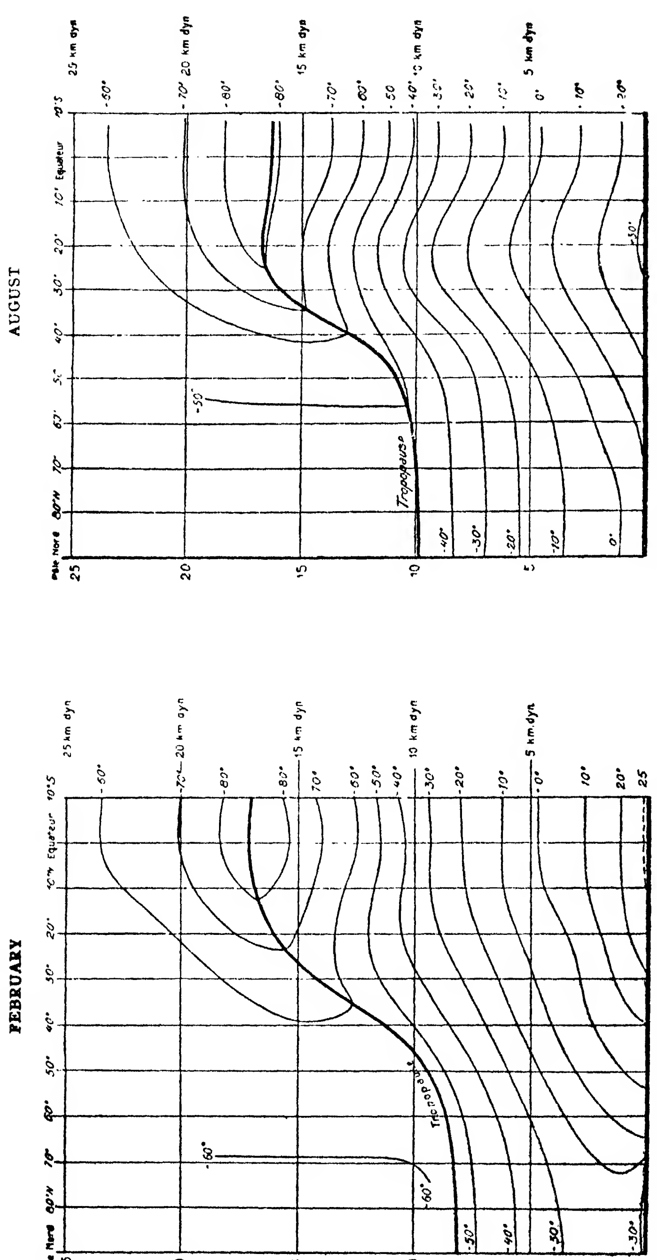

FIGURE 1.—CROSS-SECTIONS OF THE ATMOSPHERE DURING FEBRUARY AND AUGUST SHOWING THE TEMPERATURE FIELD ALONG A MERIDIAN AND THE POSITION OF THE TROPOPAUSE

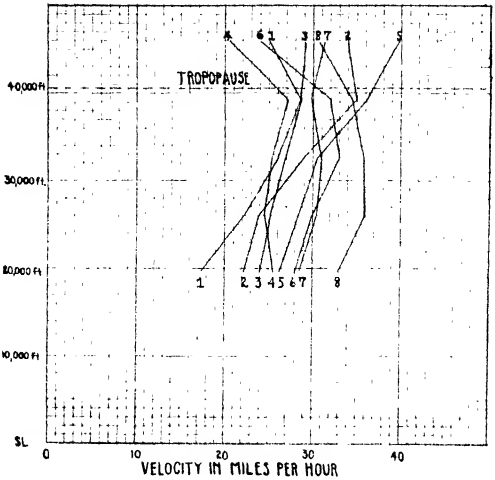

FIGURE 2.—VELOCITY DISTRIBUTION OF THE WIND AT HIGH LEVELS IN THE ATMOSPHERE OVER THE UNITED STATES

Note the steady increase in velocity up to the tropopause, then the marked decrease

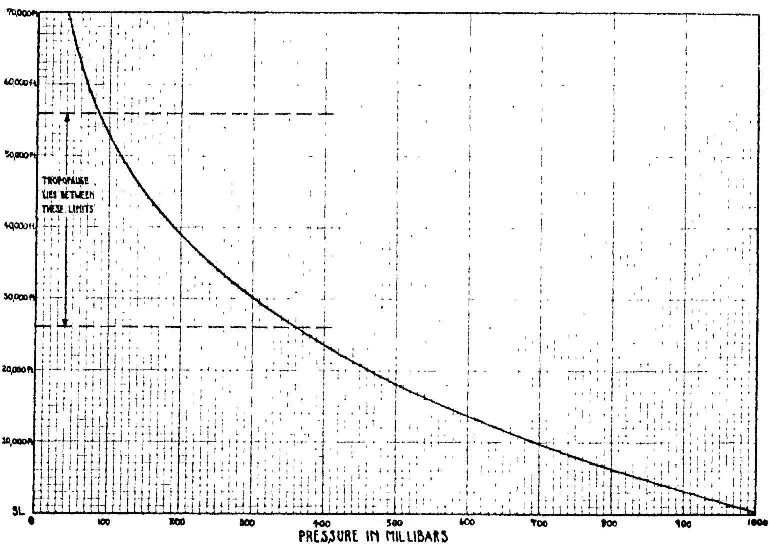

FIGURE 3.—AVERAGE ATMOSPHERIC PRESSURE AT VARIOUS ELEVATIONS

The limits of the base of the stratosphere are also shown. It may be noted that the pressure-altitude curve is very nearly a straight line below 10,000 feet. In this region 900-foot change of elevation represents approximately 1-inch change in atmospheric pressure.

show that a very pronounced rise in temperature occurs between 18 miles elevation (where it is about 223° A.), and 31 miles (where it is 343° A.).

The Ionosphere—Above about 60 miles, the number of free ions in the atmosphere increases. As a result, electrical phenomena are very important in the region between 60 miles and about 500 miles above the surface. Northern lights, or Auroras, often occur during the winter of the high latitudes in this interesting region. They have been found by Störmer and others to occur throughout the entire ionosphere and even down into the upper stratosphere.

The propagation of radio signals is greatly influenced by reflecting ionized layers in the ionosphere. The most familiar of these layers was discovered in 1902 by Kennelly and Heaviside. This layer, sometimes called the E layer, lies at a height varying between 55 and 80 miles above the surface. Above this is the F layer which is separated into the F_1 layer, lying at an elevation of about 110 miles, and the Appleton or F_2 layer at 150–250 miles. The former appears only during the summer. Recently it has been suggested that an ionized layer (the D layer) occurs in the *stratosphere* and that it tends to absorb radio waves instead of reflecting them. This may explain periodic radio "fade-outs," when all long distance communication ceases for a short time.

COMPOSITION OF THE LOWER ATMOSPHERE

The atmosphere is composed of two principal gases: nitrogen and oxygen, and a number of other minor ones. The most important of all constituents to the meteorologist is water vapor, although its total mass is relatively small. Practically all of it is concentrated in the lower 4 to 5 miles of the atmosphere. The following table, from Hann-Süring, indicates the relative concentrations of the various atmospheric gases at sea level with various concentrations of water vapor. The dryest air (0%) would never be found in nature. The air with a moisture content of 0.2% is representative of average conditions near latitude 70°; the air with 0.9% moisture represents latitude 50°; and the very moist air is characteristic of the tropics. The concentration of carbon dioxide varies greatly with most of it being found near the surface. Hydrogen is apparently not a normal constituent of the air since it readily com-

bines with oxygen. It is produced by volcanoes and its quantity is extremely variable over the earth.

TABLE I

COMPOSITION OF PURE AIR AT SEA LEVEL

CONSTITUENT	VOLUME PER CENT			
	Water Vapor Content			
	0%	0.2%	0.9%	2.6%
Nitrogen	78.08	77.9	77.4	76.05
Oxygen	20.95	20.9	20.8	20.4
Argon	0.93	0.93	0.92	0.91
Carbon Dioxide	0.03			
Hydrogen	less than 0.01			
Neon	0.0018			
Helium	0.0005			
Krypton	0.0001			
Xenon	0.000009			
Ammonia	0.0000026			
Ozone	0.000002			
Hydrogen Peroxide	0.00000004			
Iodine	trace			
Radium Emanation	trace			

COMPOSITION OF THE UPPER ATMOSPHERE

Direct observational data bearing on the composition of the atmosphere reaches well into the ionosphere. Actual samples of air have been obtained up to elevations of 20–25 miles and in this region at least, very little variation occurs in the relative percentages of the various gases (except CO_2 and H_2O of course). Above this elevation, spectral measurements of auroras have shown, that at least up to heights of 600 miles, the atmosphere still consists almost entirely of nitrogen and oxygen. Over 80 bright lines have been discovered in the auroral spectra and most of these are due to nitrogen, either as the neutral molecule N_2 or as the molecular ion N_2^+.

A bright green line at 5577 Ångström units corresponds with oxygen in the metastable atomic state.

These facts completely deny the theory, held for many years, that the various gases, above the convective troposphere, were stratified. According to this concept the gases in the higher levels arranged themselves according to their density, so that above 50–60 miles, the air consisted entirely of hydrogen and helium. No sign of these gases has ever been found in auroral spectra at any elevation, however. It now seems probable that molecular diffusion effects a complete mixing of gases at all levels, and that the composition of the atmosphere is essentially uniform at all levels (neglecting again H_2O and CO_2).

RADIATION AND INSOLATION

The principal source of energy for all atmospheric activity is supplied by radiation from the sun. Practically all meteorological elements, including wind movements, precipitation, humidity, as well as temperature are direct reflections of variations in the amount of solar energy received at various localities over the surface of the earth. Some heat energy is furnished to the earth's surface from the interior by conduction, and some through radio-active processes, but the amount supplied in this manner is entirely negligible when compared with that received from the sun.

The initial effects on the atmospheric circulation which are caused by temperature differences are of course modified greatly by the rotation of the earth, as will be brought out in the chapter on circulation, but the primary cause of all important weather events lies in variations in heat supplied to the atmosphere by the sun. The rate at which the sun furnishes heat to any point on the earth's surface is termed *insolation.*

It depends on several factors, including:

1. Inclination of the sun's rays to the plane of the horizon,
2. Distance of the sun from the earth,
3. Variations in the solar output,
4. Transmission, absorption and radiation in the atmosphere.

Inclination of the Sun's Rays to the Plane of the Horizon—The most important factor affecting insolation is the *solar altitude*,

or angle that the sun's rays make with the horizontal (figure 4). The intensity of insolation on a horizontal surface is directly proportional to the sine of this angle. It obviously varies continuously during the day as the earth revolves on its axis. It also varies at any locality during the course of a year as the tilt of the earth's axis (23.5° from a vertical to its orbital plane) causes first one hemisphere, then the other, to point toward the sun (see figure 5). When the northern hemisphere points toward the sun, as it does from the time of the Vernal Equinox (March 21) until the Autumnal Equinox (September 23), it receives a relatively large amount of insolation. When it points away from the sun during the period from September 23 to March 21, it receives a relatively small amount of radiant energy. The reverse holds true, of course, for the southern hemisphere.

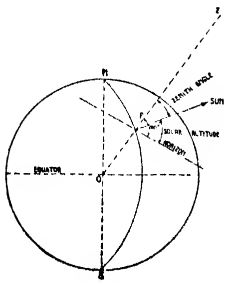

FIGURE 4.—DIAGRAM OF SOLAR ALTITUDE AND ZENITH ANGLE AT A POINT, *P*, ON THE EARTH'S SURFACE

Distance of the Earth from the Sun—Due to the elliptical path which the earth describes about the sun (which is located at one focus of the ellipse) the distance between the two bodies constantly varies. The actual deviation of the earth's orbit from a true circle is slight, however. At *perihelion*, when the earth is closest to the sun, it is about 3,000,000 miles closer than at *aphelion*, when it is most distant. Perihelion occurs on January 1, and aphelion on July 1. The average distance from the earth to the sun is about 93,000,000 miles, so the total variation amounts to but 3%. If there were no disturbing influences, and if equilibrium were reached at once, this would result in the average earth's temperature being higher by about 7° F. at perihelion than at aphelion.

Since the winters on the *northern* hemisphere occur when the earth is relatively close to the sun, they should be warmer than those on the *southern* hemisphere which occur when the earth is relatively far from the sun. Also, the northern hemisphere experiences *shorter* winters than the southern. The great difference in the distribution of land and water in the two hemispheres, however, has a much greater effect on their climate than the eccentricity of the orbit. The essential maritime character of the southern hemisphere causes it to have relatively mild winters and cool summers.

Variations in the Solar Output—C. G. Abbot has shown, through study of a long series of solar radiation measurements made at various points over the earth's surface, that the solar output is

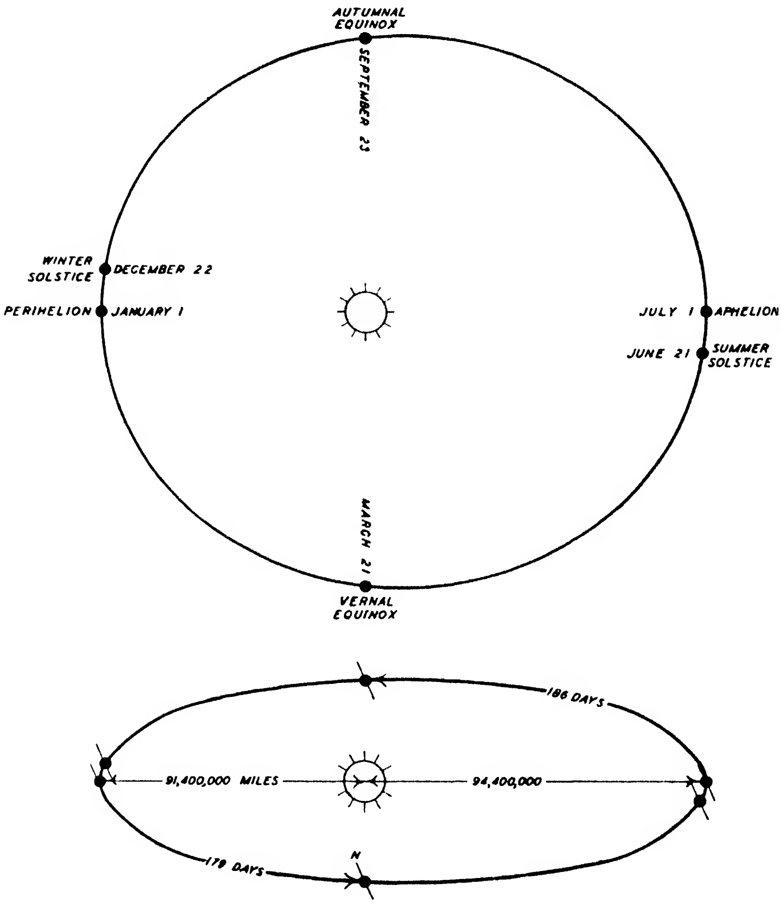

FIGURE 5.—(ABOVE) PLAN VIEW OF EARTH'S ORBIT ABOUT THE SUN. (BELOW) PERSPECTIVE VIEW OF EARTH'S ORBIT, SHOWING TILT OF THE EARTH'S AXIS

In the top figure the eccentricity of the ellipse is somewhat exaggerated.

apparently not uniform. The variations in solar energy reaching the outer atmosphere of the earth thus apparently amount to as much as 2% of the average. (The average insolation at the outer edge of the atmosphere is approximately 1.94 gram-calories per minute

per square centimeter.) Abbot has shown furthermore that these variations are definitely cyclic in nature. The sun spot cycle of about 10.5 years, as well as a number of others of smaller and greater amplitude have been recognized in the solar energy measurements.

Much research still remains to be done before it can be clearly established that there is a definite relationship between these fluctuations in solar radiation and world weather. Many factors act to mask and to confuse this relationship. With more measurements it may be possible eventually to utilize it in forecasting weather trends of medium and long periods. It seems inescapable, however, that a direct connection must exist between the amount of incident radiation and the weather on the earth.

Transmission, Absorption and Radiation in the Atmosphere—The problem of insolation is only begun with a knowledge of the solar radiation that reaches the outer atmosphere. A very complex interaction of transmission, absorption and radiation phenomena within the atmosphere itself, are involved in the *heat balance* existing between the earth and outer space. The many variable quantities make this a most difficult field to study. Particularly unfortunate is the absence of data over wide areas of the earth's surface. Certain general conclusions, however, have been reached that give a fairly satisfactory representation of average conditions.

Of the radiation which reaches the outer limits of the earth's atmosphere, approximately 42% is immediately returned to space either by direct reflection from clouds or the earth's surface (33%), or by *scattering* produced by dust particles and gas molecules (9%). Of the remaining radiation, about 15% is absorbed directly by the atmosphere, and 43% by the earth's surface. Of the 43% that reaches the earth's surface, 27% arrives directly from the sun, and 16% arrives as diffuse radiation from the sky.

Most of the radiation described in the preceding paragraph is of short wave length (light rays). Before this is returned to space it is converted to long wave heat radiation. Of the total incoming solar radiation 58% must thus be converted and returned to space (43% that reaches the earth's surface, and 15% that is directly absorbed by the atmosphere). Of the 43% reaching the surface, 8% is re-radiated directly to space and the remainder, 35%, is transmitted to the atmosphere. This 35%, together with the 15% that was directly absorbed by the atmosphere (including the clouds)

from the sun, is then re-radiated to space by the atmosphere to complete the balance between incoming short wave and outgoing long wave radiation.

Of the 35% long wave radiation transmitted to the atmosphere by the earth's surface, some is radiated *directly* to the atmosphere and some is transferred to it by the evaporation of water. Approximately 16% is radiated directly and 23% is transferred through evaporation. This totals 39%, or 4% more than the total long-wave radiation from the earth. This 4% excess is returned to the

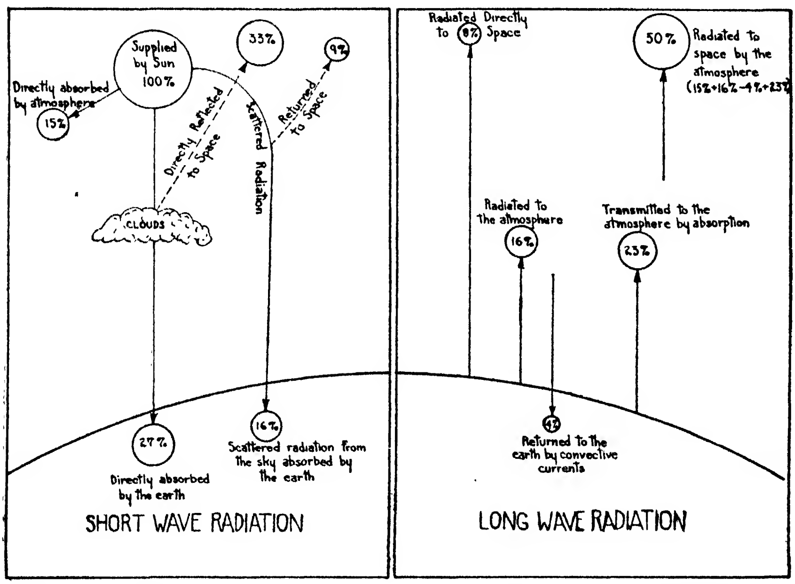

FIGURE 6.—BALANCE BETWEEN INCOMING SHORT WAVE SOLAR RADIATION AND OUTGOING LONG WAVE TERRESTRIAL RADIATION

earth by convective currents to maintain the balance between the earth and the atmosphere.

Figure 6 portrays graphically the rather complicated radiative process indicated in the preceding paragraphs.

ABSORPTION

The sun radiates as an almost perfect radiator (black body) at a temperature of about 6200° A. Figure 7 shows the distribution of solar energy at the outer limits of the atmosphere (dotted line),

and at the earth's surface (solid line). It is seen that comparatively little energy is lost in the *visible* spectrum as the sun's rays pass through the earth's atmosphere. A very large percentage of

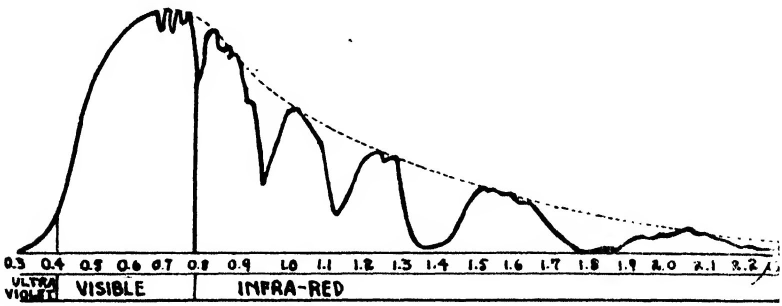

FIGURE 7.—SPECTRAL DISTRIBUTION OF SOLAR ENERGY

the *long wave* (infra-red) radiation is absorbed, however. This absorption is performed mainly by water vapor and to a minor extent by carbon dioxide and ozone. Oxygen has a single absorption band, near 0.8 μ. Other atmospheric gases are of negligible importance.

Ozone—This gas is of very great importance in its absorbing effect on *ultra-violet radiation.* W. J. Humphreys points out that if the amount of ozone in the atmosphere were decreased considerably, the ultra-violet radiation reaching the earth would be much stronger and the eyes of animals could not have developed in their present form. If the ozone concentration were much more than at present, on the other hand, animals could not maintain their health, since a definite amount of the antirachitic effect of ultra-violet radiation is absolutely essential to normal growth.

The quantity of ozone is rather slight near the surface and increases to a maximum at an elevation of 12–18 miles. Its concentration is influenced mainly by the intensity of the solar radiation. With strong radiation the amount of ultra-violet rays increases, and the amount of ozone likewise increases. Also, the ozone concentration varies with the season and with the general pressure distribution.

Smoke and dust are very important locally in decreasing the intensity of solar radiation reaching the earth's surface. At Chicago, Illinois, for example, the solar radiation during the winter months is only 55% of the value recorded at Madison, Wisconsin. Madison

is about 75 miles north of Chicago, and is by no means smoke-free itself. There must, therefore, be at least a 50% reduction in solar radiation in the Chicago region during winter due to atmospheric pollution. It is very evident that this is of considerable economic importance.

TERRESTRIAL RADIATION

The portion of the solar energy which is not absorbed by the atmosphere, or returned immediately to space by reflection and scattering, is employed in warming the earth's surface. Since the mean temperature of the earth remains practically unchanged, the earth radiates back to space all of the incident solar radiation after transforming it to long wave radiation. The average temperature of the earth may be assumed to be about 287° A. Since it radiates as a nearly perfect radiator, its curve of spectral distribution may

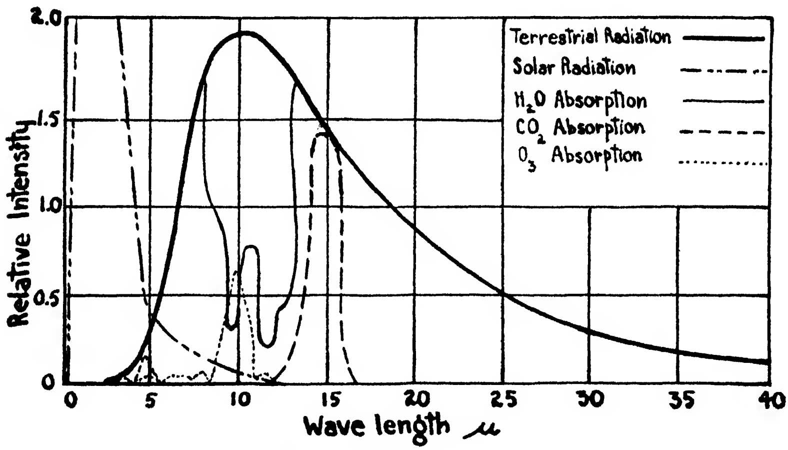

FIGURE 8.—TERRESTRIAL RADIATION AND ABSORPTION

be represented by figure 8. This curve is seen to be greatly different from the one representing the sun, since its entire radiation is in the infra-red. Also shown in figure 8 are the absorption curves for water vapor (corresponding to a column containing 3 grams of water vapor per square centimeter), for carbon dioxide (corresponding to average quantities of CO_2 at sea level), and for ozone (corresponding to a layer of O_3 about 3 mm thick at standard pressure and temperature). It may readily be seen that these gases have a very

strong absorption for the long-wave radiation from the earth. Ozone is important, since its maximum absorption occurs in a region of the spectrum where H_2O and CO_2 have little effect. Other atmospheric gases (nitrogen, oxygen, etc.) are of negligible importance in connection with long wave radiation.

These various atmospheric gases are very important in the thermal economy of the earth. Since they are relatively much more transparent to the short-wave solar radiation than to the long-wave terrestrial radiation, they exert a marked blanketing effect on the earth. It may be calculated in fact, that the average temperature of the earth would be at least 10°–15° C. lower if these gases were absent. The water vapor and carbon dioxide are concentrated in the lower levels of the atmosphere (mostly below 5 km), whereas the ozone is present high in the stratosphere. All of these gases vary greatly in their concentration from time to time and from place to place.

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CHAPTER 2

OBSERVATIONAL MATERIAL

INTRODUCTION

Accurate observation of the various weather elements is of the greatest importance for all meteorological purposes. This is especially true in the case of aeronautical meteorology where, under doubtful weather conditions, the safety of every projected flight depends on absolutely reliable observations. Furthermore, correct analysis of the synoptic weather chart is only possible when all reports are reliable and representative. Since all modern methods of weather forecasting are based upon a weather chart prepared from a number of observations all taken at the same time, it is important further that a means be available for the rapid transmission of this information to forecasting centers and airline headquarters.

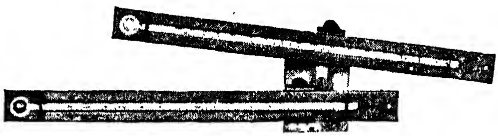
The obtaining of satisfactory data depends upon selecting personnel who fully understand the making of weather observations and providing them with accurate instruments to indicate the weather elements. The dissemination of this information requires a closely coordinated transmission network including the facilities of radio, telegraph and telephone. In this chapter methods of obtaining satisfactory observations will be considered, followed by a description of the communications systems employed in various countries for disseminating this information.

WEATHER ELEMENTS

A number of elements within the atmosphere may be measured, each one of which may be of use to different interests and at different times. Fundamental elements of interest to all meteorologists are *air temperature, atmospheric pressure, wind direction and velocity, present state of the weather* and *moisture content of the air*, as measured by its humidity. The *rate of change of barometric pressure* is of considerable importance to forecasters. The *amount of*

precipitation is of general interest, but is of especial importance mainly to agriculturists and to those making general climatological studies. The amount of detail required in reporting the various elements varies greatly. To some interests the *state of the weather* may be expressed with ample detail as *clear, cloudy, raining* or *snowing*, whereas the airplane pilot may wish to know how much sky is covered by clouds, how high the clouds are, how intense the precipitation is, etc. All users of weather information are greatly interested in having the individual reports as representative of their general region as possible.

Temperature—It has been found that the obtaining of thoroughly representative temperature measurements is a very difficult problem due to the wide local variations in this element. The thermometer which measures the temperature of the air must not be exposed too near to the ground so that insolational heating during



Courtesy Taylor Instrument Company

FIGURE 9.—MAXIMUM AND MINIMUM THERMOMETERS—STANDARD U. S. WEATHER BUREAU MODEL

the day or radiational cooling at night may give rise to abnormally high or low temperatures. The thermometer must be shielded from the direct rays of the sun and it must be properly ventilated so that it will indicate the true air temperature unaffected by direct solar radiation. Various types of thermometers are commonly used, including directly indicating types such as the mercury (figure 9) and the alcohol thermometer, indirectly indicating types which depend on unequal expansion of two metals, or between a metal and a liquid as used in various types of thermographs, the variation of electrical resistance of a solid or a liquid, or of the electromotive force set up at a junction between dissimilar metals. For obtaining single observations the ordinary mercury or alcohol thermometer is accurate and entirely suitable. To obtain a continuous record of temperature, the thermograph is used in which the unequal expansion of a solid and a liquid is indicated by a pen resting on a revolving drum. The accuracy of this method of measuring temperature generally does not equal the directly indicating methods, since friction

in the recording mechanism and fatigue in the metals used in the expansion elements cause a slight lag and very slight drift, necessitating occasional recalibration. By careful design these errors may be practically eliminated, however, and for all practical purposes the thermograph is entirely satisfactory. Remote indicating devices employing thermojunctions are very convenient and entirely reliable if occasionally recalibrated.

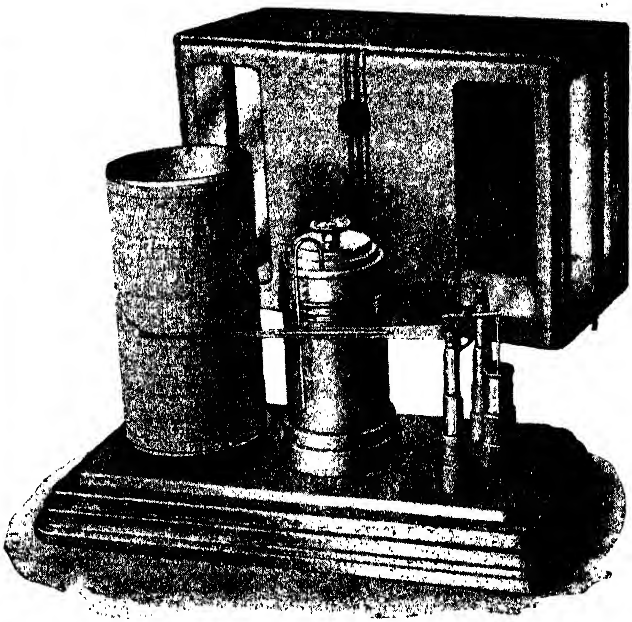
The importance of obtaining temperatures which are truly representative of a given area can hardly be overestimated since the number of available reports is always very limited and even with a fairly close network of observation stations, each one must serve for several thousand square miles. Incorrect exposure of the thermometer may give consistently high, or consistently low, or simply irregular values of the temperature. It is generally found that the temperatures obtained at weather observatories located in large cities are considerably higher than the average for the region and for this reason there is some advantage in having all weather observations used in forecasting taken at the airports, which are generally located several miles from the center of town. Even here care must be taken in the location of the instrument, for it will be observed frequently, particularly in the summer-time, that the air temperature will be consistently too low if it is taken in the middle of large green fields or will be consistently too high if taken near large concrete areas, such as runways. It is generally possible to avoid these local effects by locating the thermometer 20 to 30 feet above the surface of the ground on the roof of a building, so that the effects of the air immediately adjacent to the surface are eliminated. For the best results the thermometer should be located at least 10 feet from the roof of the building, but this is not always possible.

Pressure—The accurate determination of the atmospheric pressure at any locality is generally much simpler than that of the temperature. Purely local variations do not occur so that the location of the recording instrument is of little importance and representative values are readily obtained. Pressure measuring instruments are of two general types—the mercurial and the aneroid. The mercurial barometer measures the pressure of the atmosphere indirectly by employing a column of mercury to balance the column of air located above it in the atmosphere. This is also the most accurate means of measuring the pressure since no mechanical devices are required to transform the pressure indications obtained

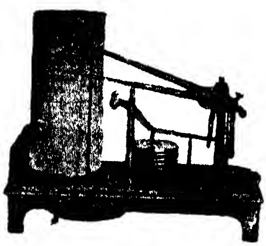
in this manner into a usable form. Mercury is used in this type of barometer since it has a very high density, and only a comparatively short column (approximately thirty inches at sea level) is required to balance the weight of the entire column of the atmosphere above it. At higher elevations the air column above the barometer becomes less and less and the length of the mercury column required to effect a balance becomes shorter and shorter. This decrease in the length of the mercury column amounts to roughly one inch of mercury for each 900 feet of vertical ascent within the lower few thousand feet of the atmosphere.

Since the mercurial barometer (figure 10-a) provides a simple and reliable means of measuring atmosphere pressure, and since other pressure measuring instruments are generally referred to it, most pressure measurements are given in terms of the height of the mercury column necessary to balance the atmospheric pressure at any point. Thus, it is common to indicate the atmospheric pressure as being, "so many inches (or millimeters) of mercury" rather than as a pressure of "so many pounds per square foot, or dynes per square centimeter." Another system of pressure measurement that is supplanting all others gives the atmospheric pressure in terms of a standard pressure called a *bar*. With this system of denoting atmospheric pressure, normal sea level pressure is 1 bar or 1000 *millibars*. The pressure at other times and localities may then be expressed directly in *millibars*. Any of the various systems of indicating pressure mentioned above may be converted into other systems by means of simple tables. The unit *millibar* is fast becoming the standard pressure unit for much of the world; the unit *inch of mercury* is still employed rather widely, as is the unit *millimeter of mercury*. Generally the meteorologist is interested in variations of pressure amounting to approximately .01 of an inch of mercury. To attain this degree of accuracy with the metric or the millibar system, it is necessary to give the pressure to within two or three tenths of a millimeter or of a millibar.

Although the mercurial barometer is perhaps the most accurate means of determining the pressure, it is not especially convenient to use because of its bulk and, furthermore it is not readily possible to obtain continuous records from it. For general purposes, therefore, the *aneroid* barometer is widely used. This instrument depends for its operation upon the change in shape of elastic metal cells. These are made in the form of flattened conical discs joined



a *b*



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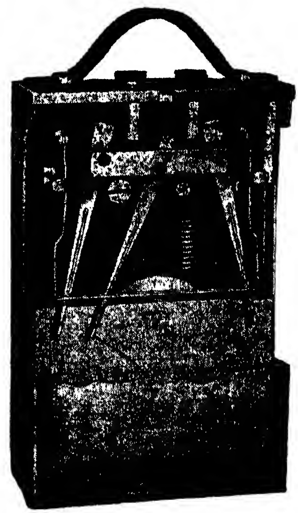


d

FIGURE 10.—(*a*) MERCURIAL BAROMETER. (*b*) MICROBAROGRAPH. (*c*) DUAL TRAVERSE BAROGRAPH. (*d*) AIRCRAFT FLIGHT ANALYZER

Courtesy Julien P. Friez & Sons, Inc.

The microbarograph is used to record small pressure variations on an open scale. The dual traverse barograph is used to record high altitude flights on an open scale. The aircraft flight analyzer is placed in transport airplanes to furnish a continuous record of altitude maintained. Auxiliary pens also record operation of the radio transmitter and the automatic pilot.

together and partially evacuated. As the pressure surrounding one of these cells increases or decreases its shape will change. This may be magnified by means of a mechanical system and either registered on a dial by means of a pointer or recorded on a strip of paper by means of a pen (figure 10-b, -c, -d). In order to obtain a relatively open scale, it is generally found desirable to employ several of the evacuated cells so that the changes in shape of each cell will be added to those of the others. Practically all portable pressure measuring instruments are of this general type, including the small portable *aneroid barometer*, the *barograph* used for making continuous records of atmospheric pressure, and the sensitive *altimeter* used in airplanes for registering altitude. Due to the elastic stresses present in the metal of which the evacuated cells or *sylphon* is constructed and also due to friction in the magnifying and registering mechanism, the aneroid barometer is inherently subject to several types of error: (1) that which appears as a gradual but continuous shift in the pressure indication due generally to elastic stresses; (2) that which is evidenced by minor irregularities, due generally to friction in the mechanism; (3) that which is due to lag of the instrument in cases of rapid changes in pressure. The first type of error may be eliminated by allowing ample time for the instrument to become stabilized after being moved. The second type of error may be eliminated by frequent comparison of the aneroid barometer with a mercurial type of instrument, and the third can be eliminated only by very careful construction to eliminate friction in the registering mechanism and within the sylphon element itself.

Pressure Reduction—In order to construct charts of the atmospheric pressure over wide regions, portions of which lie at various elevations above sea level, it is necessary to adopt some means of reducing the observed atmospheric pressure to a standard level. Without such a procedure it would be impossible to construct isobaric charts over the surface of the land, for, as has been pointed out above, the variation in atmospheric pressure due to differences in elevation amounts to approximately 1 inch of mercury per 900 feet of elevation. Since ordinary atmospheric pressure changes connected with the passage of storms commonly do not exceed several tenths of an inch, it is clear that such changes would be entirely masked by differences in elevation if the pressure were not reduced to a common level. This process is comparatively simple near the shore. Here a mean value of the temperature of the hypothetical

air column lying between the barometer and the surface of the ocean can be assumed and from this the weight of the imaginary air column can be readily determined. This is illustrated in figure 11 in which the actual atmospheric pressure at station A, at an elevation, d, above sea level is known. This pressure may be changed so that it will represent the pressure which A would experience if it were lowered to sea level at B by removing a section of the earth's crust. This can be done if the distance, d, is known and the temperature of the column of air from A to B is known or inferred. It is, of course, impossible to know what this temperature would be but a fair approximation can be obtained by utilizing the average temperature within the preceding twelve hours observed at station A. This is the method commonly used over most regions in *reducing* observed atmospheric pressure to sea level values. It is an artificial

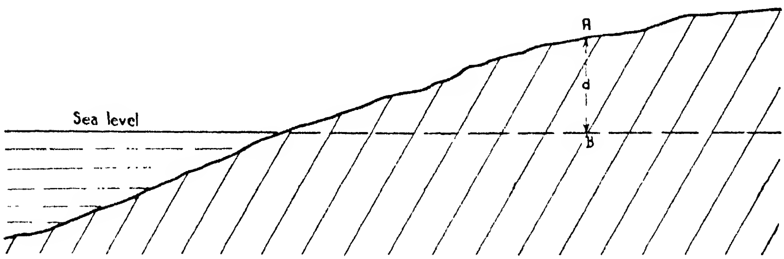


FIGURE 11.—BAROMETRIC REDUCTION TO SEA LEVEL

method in some respects but if all stations in a given area reduce their pressures in the same manner the resulting values will be at least comparable.

While this method is quite satisfactory in regions within a few hundred miles of the coast and at moderate elevations, it is found necessary to modify it somewhat in reducing observations made at inland and high level stations. The methods of making such reductions, the so-called "plateau reduction," have been extensively studied by several investigators as a result of which rather complicated formulae have been developed for use under such circumstances. These take into account the average barometric pressure of the region, its average temperature, the differences between its average pressure and that of surrounding stations and include empirical corrections which are determined for each case through experience. With all the effort that has been expended on methods of determin-

ing pressure reductions in plateau areas, this subject is still in a very unsatisfactory state and pressure observations obtained from high level inland stations are subject to great errors when reduced to sea level. This is due to the large percentage of the total sea level pressure which must be added to the station pressure. A station at 5000 feet elevation thus may have a station pressure of only 25 inches. To this must be added approximately 5 inches to reduce it to sea level. Any error involved in determining the temperature of the hypothetical air column between the station and sea level will cause a very large error in the sea level pressure. It is quite obvious that in inland regions, under conditions where the temperature at the station may be widely different from the normal value, errors of this type may be considerable. It has been suggested that, in order to eliminate such misleading pressures as are obtained under these conditions, it might be desirable to express the atmospheric pressure in *deviation from a normal value* rather than in an actual sea level reading. It would still be possible to draw isolines in this manner, connecting points of *equal departure from the mean* in the same manner that isobars are drawn connecting equal values of the sea level pressure. This method has not been worked out practically, but it offers some promise as a means of eliminating the errors which are inherent in the present method employed in obtaining sea level pressures.

The question of obtaining representative pressures is generally of little importance since the variations in pressure from place to place over the surface of the earth generally occur smoothly. The only precautions that should be taken in this regard are to expose the barometer to the true pressure of the outside atmosphere. This matter may assume some importance in modern air conditioned buildings, where at times the pressure difference between the interior and exterior of the building may amount to several hundredths of an inch. This holds true especially for altimeters on airplanes. In this case the pressure within the cockpit may differ from the outside pressure by several hundredths of an inch, which might amount to an altitude difference of 50 to 100 feet in some circumstances.

ALTIMETERS

Instruments used for determining altitude by means of atmospheric pressure are all variations of the aneroid barometer (see

figure 13-b). In constructing the indicating mechanism of such instruments it is commonly assumed that the temperature decrease within the atmosphere occurs at a uniform rate of 3.57° F. per 1000 feet. This value of the temperature gradient (NACA standard atmosphere), which is an average for the actually observed gradient at 40° latitude, is, of course, arbitrary, and will only by chance represent the actual conditions prevailing at a particular time. As a result, *pressure readings from such an altimeter cannot be compared directly with the pressure determined by the method of reduction to sea level as described above in which the actual air temperature is used in making the reduction.* At times the pressures may agree, but ordinarily there is a difference amounting to at least several hundredths of an inch and in extreme cases the pressure measured by a sensitive altimeter, as used in airplanes, may differ from that determined by reducing the actual atmospheric pressure as measured by a mercurial or aneroid barometer to sea level, by as much as two tenths of an inch. Thus, if the movable scale on a sensitive altimeter is set to the barometric pressure obtained by reduction of the actual station pressure to sea level, using the commonly employed formulae, the altitude indicated on the sensitive altimeter may be in error by as much as 150 to 200 feet. Since this is an error of considerable magnitude, it is important that sensitive altimeters in planes in flight be set to coincide with the pressure indicated by other altimeters *of the same type* at ground stations, rather than with the pressure obtained in reducing actual atmospheric pressure to sea level, using the ordinary formulae. In this manner the error due to different methods of reduction disappears, since both the instrument at the ground station and the one in the airplane employ the same methods of reduction. It is possible to obtain a *standard atmosphere sea level pressure* from any barometer by the use of appropriate formulae. Such a pressure may then be used for setting altimeters.

Humidity—The amount of moisture present in the air is termed the *humidity.* The higher the temperature, the larger the amount of moisture which may be held in the air without reaching saturation. When a particle of unsaturated air is cooled it will approach nearer and nearer to the *dew point.* When this point is reached the air cannot hold all of the water present and it is then precipitated as liquid or solid water in the form of dew or frost. The humidity may be reported in several ways, each of which has

its special advantages. The dew point has already been mentioned and is a valuable way of expressing humidity since it shows at once, if the temperature is known, the amount of cooling necessary to reach saturation. The degree of saturation is also given by the *relative humidity*, which expresses it as a percentage, with 100% indicating completely saturated air and 0% perfectly dry air. The relative humidity, unlike the dew point, varies with the temperature of the air and thus will vary at different points in the same body of air as the temperature changes. The *absolute humidity* expresses the *weight* of water in a given volume of air. Since the volume of a given weight of air is dependent on the pressure, it is evident that this expression for the humidity will change as the atmospheric pressure varies. The *vapor pressure* gives the *partial pressure* of the water vapor. The most satisfactory of all means of reporting humidity, at least for many meteorological purposes, is the *specific humidity*. This measures the *weight* of water contained in a given *weight* of air and it is, therefore, independent of temperature and pressure changes as long as no condensation or evaporation occurs. The specific humidity is generally used in meteorological calculations which are concerned with comparing the properties of different types of air, since it is not affected by ordinary changes as the air is raised or lowered in the atmosphere. The use of specific humidity is described in some detail in chapter 3. The humidity of the same sample of air at a temperature of 59° F. and a pressure of 29.92 inches of mercury with a specific humidity of 6.8 grams per kilogram may thus be given as:

Specific Humidity	6.8 grams per kilogram.
Dew Point	47°
Relative Humidity	64%
Absolute Humidity	8.22 grams per cubic meter.
Vapor Pressure	.322 inch.

The interrelationship of these various expressions for humidity will be discussed in chapter 3.

Humidity is generally measured by determining the rate of evaporation of water into the air. This is generally accomplished by the use of a *psychrometer*, an instrument consisting of two thermometers, the bulb of one being surrounded by a moistened cloth from which water is free to evaporate and the other being freely

exposed to the air (figure 12). Due to evaporation, the temperature indicated by the thermometer with the "wet-bulb" will be somewhat lower than the other, the more rapid the evaporation the greater the difference between the two thermometers. When the air is saturated the two thermometers will indicate the same temperature, while if it is very dry the wet-bulb thermometer may indicate a temperature many degrees lower than the other. By the use of suitable tables this difference in temperature may be used in calculating the humidity in any of the forms indicated above (see References). In using such an instrument it is important that the amount of ventilation be adequate to insure that the wet-bulb thermometer will indicate the lowest possible temperature under the conditions prevailing. It is common practice to employ a fan to cause rapid circulation of the air about the thermometer bulb, or to swing the thermometer rapidly through the air. Another device very commonly used for measuring humidity depends for its operation upon the fact that a strand of hair will change its length in atmospheres of different humidities. The *hair hygrometer* consists of several strands of stretched human hair which are arranged so that any change in their length is indicated by a pointer moving over a dial. This instrument is fairly accurate although it must be recalibrated frequently. It is very convenient since it may be arranged to read directly in relative humidity. It also has the advantages of light weight, and of being easily arranged to give a permanent record of humidity changes on a strip of moving paper.

Courtesy Taylor Instrument Company

FIGURE 12.—SLING PSYCHROMETER

In a third type of humidity measuring device a polished metal cylinder is filled with a liquid such as alcohol or ether whose evaporation will lower the temperature of the metal cylinder. By blowing air through the liquid, the temperature may be lowered until dew

begins to form on the polished metal cylinder. By this means the dew point may be directly determined.

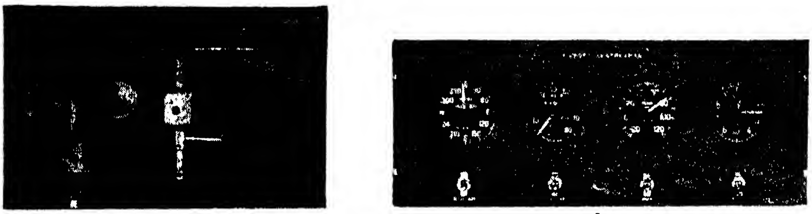
When the temperature is much below freezing, all of the above methods of measuring humidity are rather unsatisfactory and they become almost completely unreliable at temperatures below $-10°$ F. It is thus very difficult to obtain satisfactory humidity measurements in extremely low temperatures, and it is believed that an entirely new line of attack on this problem will be necessary before successful solution is obtained. It has been suggested that since the transmission of long wave lengths of light is directly related to the humidity, that it might be possible to construct a humidity measuring instrument based on this principle which would measure the amount of transmission of infra-red light.

Wind—As might be expected, the obtaining of representative values of wind direction and velocity is a difficult matter. Wind near the surface is affected by many local influences, as is the case with temperature and in fact the same influences frequently affect both elements. Irregularities in topography cause corresponding variations in both direction and velocity of the wind near the surface. In mountainous regions these variations may extend to heights of several thousand feet above the ground. Bodies of water, by their tendency toward equalizing temperatures, also control the wind in their vicinity to considerable extent, thus the well known *sea breeze* and *lake breeze*. Even very local influences affect the wind, especially when it is light, thus there is a tendency for a light wind to blow from green fields to adjacent uncultivated land during hot afternoons. In thickly settled regions large buildings affect the surface wind considerably. It is thus seen that thoroughly representative wind reports are difficult, if not impossible, to obtain. Precautions ordinarily taken to insure satisfactory wind information include some means to eliminate purely local effects due to minor irregularities of the ground and the presence of buildings. The *wind vane* used to indicate the wind direction and the *anemometer* used to show wind velocity are thus located a considerable distance above the ground and also above surrounding buildings, to eliminate surface gustiness. Whenever possible the station at which wind measurements are taken should not be located too close to hills and valleys, which generally produce a persistent wind from a certain direction. Measurements taken at the top of hills are also un-

reliable, since the wind velocities in such places are generally considerably higher than the average for the region.

Although the velocity and direction of the wind are considerably affected by local influences, nevertheless, wind information is of great use to the meteorologist if proper allowance be made for possible errors. Deviations of the observed winds from the average regional values disappear to a large extent as the general circulation increases and as the weather situation becomes well developed.

The wind direction is measured by a *wind vane.* This is commonly arranged to indicate the direction at some distance by means of an electrical circuit. In the simplest type of transmitting wind vane, the direction is given to four points of the compass by means



a *b*

Courtesy Julien P. Friez & Sons, Inc.

FIGURE 13.—WIND DIRECTION AND VELOCITY INDICATORS

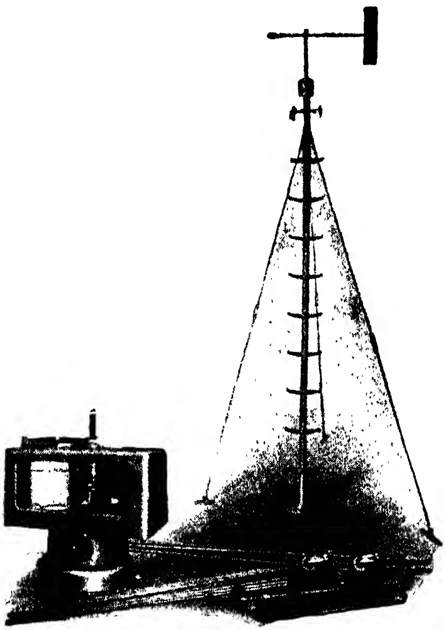
These instruments are of the transmitting type. Self-synchronous generators at the base of the wind vane and anemometer (*a*) transmit the measurements of wind direction and velocity to an indicating unit (*b*). This indicating unit may be located a considerable distance from the recorders. This unit shown also includes an outside air thermometer and a sensitive altimeter.

of a simple commutator mounted on the wind vane. In more elaborate installations the direction may be indicated very accurately by means of two self-synchronous motors, one mounted at the wind vane and one at the recording position (see figure 13).

The wind velocity may be determined by observing the speed of rotation of a ***multiple cup anemometer*** (see figure 13), by measuring the pressure on a flat surface, or by using a ***pitot static tube*** (see figure 14).

The cup anemometer is one of the most commonly employed methods for measuring surface wind velocity. It is especially desirable for this purpose since it is very accurate in indicating low velocities. With this instrument the velocity of rotation of the cups is measured by noting the number of revolutions in a given interval

of time. From this the wind velocity may be determined from the calibration chart of the instrument. The cup anemometer is more or less of an integrating device and thus gives reliable average measurements of the wind when its velocity is changing rapidly. It is not adapted to measuring sudden changes in the velocity, however,



Courtesy Julien P. Friez & Sons, Inc.

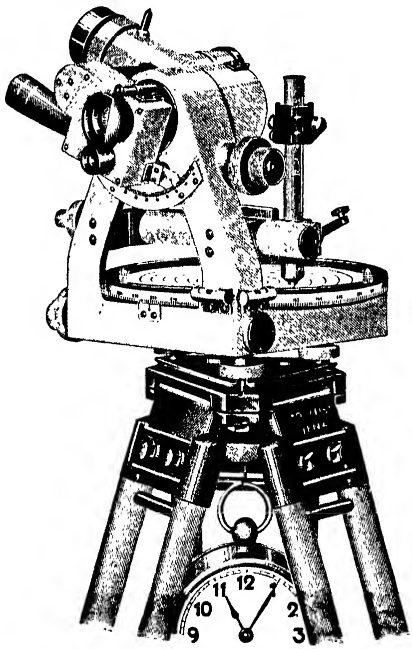
FIGURE 14.—ANEMOGRAPH—MECHANICAL WIND DIRECTION AND VELOCITY RECORDER

This instrument operates on the pressure tube principle, instead of employing rotating cups. The wind vane and pressure tube unit is directly connected to the continuous recorder below. This type of velocity indicator is desirable for recording gustiness, since it is instantaneous in operation. It is also useful in regions affected by hurricanes, since it is not affected by power failure.

and for this purpose some form of a pitot-static opening is employed. This type of instrument measures the difference in pressure between two openings, a dynamic one facing the wind and a static one at right angles to the wind direction. The difference in pressure of the two openings can be converted into wind velocity.

Upper Wind Measurements—To measure the wind velocity at any distance above the surface requires some form of triangula-

tion. If the height of clouds is known it is possible to determine their velocity by noting their rate of angular movement from the surface. Commonly, however, upper air observations are made by releasing a small balloon and following its path of ascent by means of a theodolite (figure 15). Measurements are taken both of the vertical and horizontal angles to the balloon at fixed intervals during its



Courtesy Carl Zeiss Inc.

FIGURE 15.—PILOT BALLOON THEODOLITE

This instrument records the horizontal projection of the flight of the pilot balloon directly on a card attached to the theodolite, as the flight is in progress. The calculation of the data is thus greatly simplified.

ascent by one or two theodolites and from the information thus obtained it is possible to plot the horizontal projection of the path of the balloon. From this curve it is then possible to determine the rate of horizontal movement of the balloon at any time in its flight and thus to determine the wind velocity. If a uniform rate of ascension for the balloon be assumed it is possible to obtain all necessary information from one theodolite. In actual practice it has been

found that the results attained from the use of a single theodolite, assuming a constant rate of ascent, are nearly as reliable as those using two theodolites, since the vertical currents which influence the rate of ascent are generally negligible. Occasionally, at times of marked convective activity the actual rate of ascent may vary somewhat from the assumed value and give rise to incorrect determinations of upper wind velocities.

Observations may be carried on in this manner only up to the base of the lowest cloud layer and, therefore, at many times when upper wind information would be of the greatest value, as during general storm conditions, it is not possible to obtain this information. Experiments are being carried on at the present time with balloons carrying small radio transmitting sets which are followed from the ground by directive receivers so that observations may be obtained even when visibility at the ground is zero, and it is expected that developments along this line will make possible the regular obtaining of upper wind observations in all types of weather within the very near future.

Ceiling—The term ceiling as employed in aviation is defined as the distance between the surface of the ground and the base of the lowest cloud layer. Some lack of uniformity is present in the use of the word ceiling, however. In some cases it is used in the strict sense given above, and in others it is taken to mean the elevation of the lower portion of the clouds above *sea level.* This confusion in methods of reporting the term *ceiling* is undesirable and may result in misleading information at times. It is believed very desirable to limit the use of the term *ceiling* to the strict usage defined above, that is, to the *distance between the ground and the base of the lowest broken or overcast cloud layer.*

Where it is necessary to report the base of the cloud layer above sea level, in cases where the elevation of the ground is not known, as frequently occurs with pilots flying over irregular topography, expressions similar to the following should be employed, "base of clouds 2000 feet above sea level" or "entered clouds at 4000 feet above sea level." There will thus be no confusion with the strict usage of the term *ceiling*, and other persons employing this information will not be misled. Figure 16 indicates the manner in which the *ceiling* differs from the *elevation of the cloud base* over rugged topography. At sea level, the two qualities are equal (a). At points B and C, the ceiling (b and c) is less than the elevation of

the cloud base (a). At point D the ceiling is zero, yet the elevation of the cloud base is still the same as before (a).

The ceiling height may be determined in a number of different ways. Perhaps the most satisfactory of these is the determination by an airplane which is flying at the base of the cloud layer, which can thus measure its elevation without any question. The pilot can in this manner clarify cases which are confusing to an observer from the ground, particularly those in which an isolated cloud over a ground station may give a false idea of conditions in the immediate vicinity. This method of ceiling determination is only practicable,

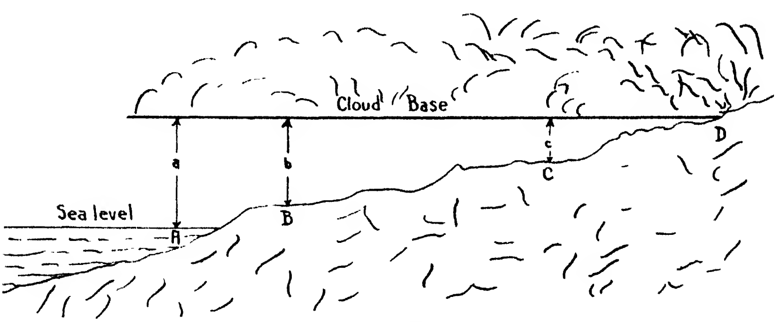


FIGURE 16.—EFFECT OF TOPOGRAPHY ON CEILING HEIGHT

of course, when an airplane happens to be flying in the vicinity of a station desiring a ceiling report.

Three methods are commonly employed by ground observers in measuring ceiling height. With the first method the observer on the ground notes the intersection of clouds with prominent objects whose elevations are known. In flat country very low ceilings may thus be determined by the intersection of cloud layers with tall buildings or smoke stacks, and in mountainous country the point at which clouds intersect known elevations on surrounding hills may be noted. At times when this method of intersection with known objects is not practicable due to poor visibility or lack of sufficiently tall objects, the ceiling may be determined by releasing a small balloon inflated to a known *lift* so that it has a known rate of ascent. By noting the time required by the balloon to enter the clouds after leaving the ground, it is then a simple matter to determine the distance between the ground and the cloud base. This method of ceiling determination is unreliable whenever there is precipitation, how-

ever, for the weight of the rain or snow changes the ascension rate sufficiently to invalidate measurements made under these conditions. At night, determination of the ceiling height is not generally possible by either of the methods just mentioned and a still different method must be used. The point at which a vertically directed high intensity beam of light intersects the cloud base is observed with a clinometer. If the horizontal distance of the observing point from the beam is known, as well as the vertical angle from the observing point to the intersection of the beam with the clouds, it is then a simple matter to calculate the ceiling height by the use of trigonometry. This method of observing the ceiling is not particularly successful when the air near the surface is polluted with smoke and dust, for it is then impossible to obtain a sharp intersection of the beam with the cloud base. It also fails during heavy precipitation, for the falling rain or snow scatters the light beam in many cases before it reaches the cloud base.

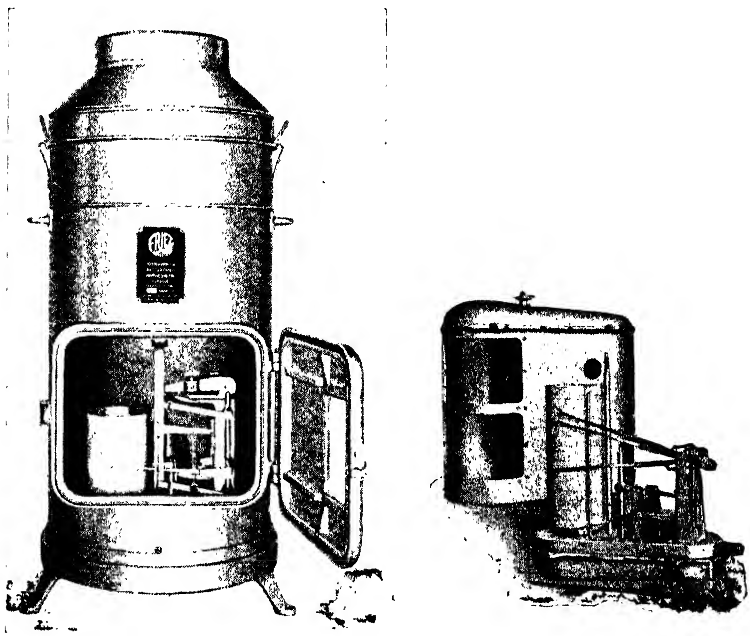
It may be seen from the above discussion that all of the present means of ceiling determination are subject to considerable error under certain circumstances. For this reason ceiling reports under the more difficult situations mentioned above must be accepted with considerable reservation.

Visibility—Visibility is defined as the horizontal distance at the surface at which objects may be seen by the unaided eye. The visibility may vary considerably within the lowermost layers of the atmosphere, being generally poorest at the ground. Ordinarily only the surface visibility is reported since this is most important to the pilot making a landing or takeoff. At the present time, visibility is measured simply by noting the distance at which well known objects may be seen. Since this is actually the way the information is used by pilots, it is perhaps the most satisfactory way of indicating visibility, although other more precise methods of measuring it have been suggested.

It is important during the daytime to observe objects of approximately the same degree of contrast compared with their background, in order to make observations in various directions and at various times comparable. Similarly at night, when lights are generally observed to determine visibility, it is important that they be of approximately the same intensity. It is always desirable to report the cause of poor visibility, since it is of considerable importance to know whether the restricted visibility is due to a purely

local effect such as smoke or whether it is caused by a widespread phenomenon such as fog or rain.

Rainfall—Although rainfall measurements are of little interest to flying personnel, they are important to the aeronautical meteorologist and to some extent to dispatching personnel. Rainfall is measured either in a simple cylindrical rain gauge where the amount of water falling in a given time is measured with a graduated stick, or



a *b*

Courtesy Julien P. Friez & Sons, Inc.

FIGURE 17.—(*a*) CONTINUOUS WEIGHING RAIN GAUGE. (*b*) AEROMETEOROGRAPH

in a *tipping bucket* rain gauge which makes a continuous record of rainfall on a chart, or in a continuous weighing gauge (figure 17-a).

Aerological Soundings—Measurement of temperature, pressure and humidity by various types of meteorographs is widely employed for determining the distribution of these elements in the upper levels of the atmosphere. The meteorograph consists of a drum rotated by clockwork which carries on it a sheet of paper or smoked metal. Several pens or styluses are arranged to draw traces on this rotating cylinder, indicating the temperature, pressure and hu-

midity as continuous lines. The temperature is measured by a simple bimetallic thermometer, the pressure by an aneroid barometer and the humidity by a hair hygrometer (figure 17-b).

The meteorograph must be frequently calibrated under standard conditions to render its readings reliable. This is ordinarily accomplished by placing the instrument in a chamber connected to a mercury manometer and noting the readings of the instrument under various pressure conditions. The same procedure is followed for different temperatures and humidities.

The meteorograph is generally attached to an airplane by a shockproof mounting and carried aloft at a slow rate of ascent. This is necessary in order to allow the meteorograph to reach equilibrium with its surroundings as conditions change, since the various elements of the instrument are generally subject to more or less lag.

Radio Meteorographs—Much interest has been shown recently in developing meteorographs which are capable of transmitting by radio a record of conditions which they encounter as they ascend into the atmosphere attached to small free balloons. A number of such types have been developed within the last few years, some of which have yielded very satisfactory results. This method of obtaining upper air information promises to supplant the use of airplanes in the very near future. Aerographic ascents made by airplanes are rather unsatisfactory since they are greatly dependent on the weather conditions. Thus at times when information is most desired about the conditions of the upper atmosphere, during severe storms, it is the least likely to be obtainable. The use of the radio meteorographs will eliminate this difficulty and allow the obtaining of upper air information in all weather conditions.

Several types of radio meteorographs have been developed. Most of them include a small battery operated transmitting set which transmits on a very high frequency (from 30 to 120 megacycles). (See figure 18.)

The elements of temperature, pressure and humidity are generally measured by a bimetallic thermometer, a single cell aneroid barometer and a hair hygrometer, although one promising type employs a cell of sulfuric acid whose varying electrical resistance measures the temperature, and an unglazed porcelain tube which measures the humidity by measuring the electrical resistance of the surface of the tube.

In most types of radio meteorographs a small clockwork mecha-

nism is employed to provide a constant reference point for the meteorological elements. It is generally arranged so that the hand

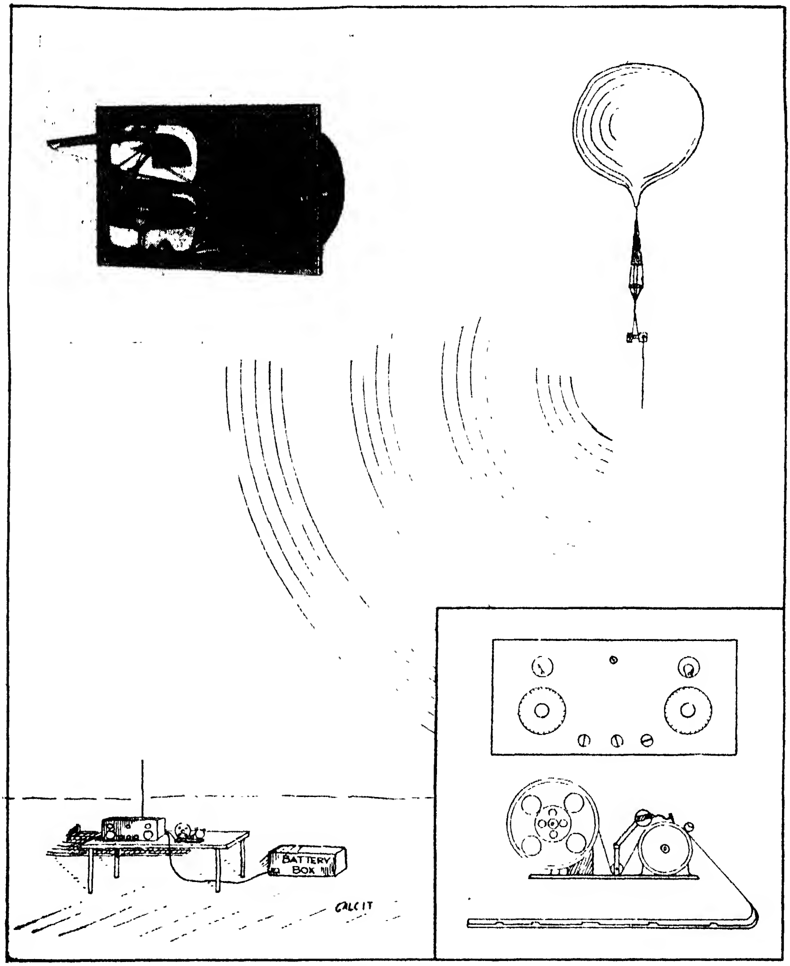


FIGURE 18.—RADIO METEOROGRAPH

Model developed by California Institute of Technology. It operates on a very high frequency (over 100 megacycles). The entire instrument as sent aloft, including wet batteries, weighs less than one pound. The radio signals are received at the ground by a simple super-regenerative receiver, and recorded on a moving tape, as shown schematically in the lower right of the figure. Remarkable accuracy and reliability have been attained with this instrument.

of a small watch rotates once every 30 or 60 seconds and in doing so makes contact with a reference point and with each of the three

elements being measured, temperature, pressure and humidity, during each revolution of the hand. At each of these contacts, the signal of the radio transmitting set is either interrupted briefly or is transmitted briefly. The reference point for the rotation of the hand is distinguished by being somewhat longer than the signals representing the temperature, pressure and humidity. Each of the weather elements is then recorded within a certain time interval following the reference point. It is thus possible to determine the actual values of these elements by noting the time which elapses between the reference signal and the signals denoting each of these elements. Another instrument employs the variations in the pressure to change the amount of resistance in an electrical circuit and to vary the type of signal emitted. This instrument dispenses with the clockwork mechanism.

The signals transmitted by the radio meteorograph are received on the ground and recorded on a revolving drum. Various means are employed to interpret the various signals transmitted and thus to determine the actual values of the temperature, pressure and humidity at various altitudes.

AIRWAYS INSTRUMENTAL EQUIPMENT

It has been found from considerable experience that certain weather elements are of especial importance to airline meteorology and that accurate measurement of these elements is essential. Each of the following weather conditions must be observed and instruments provided for their observation when necessary: 1. Ceiling, 2. Sky condition, 3. Visibility, 4. Obstructions to visibility, 5. Precipitation, 6. Temperature, 7. Dew point, 8. Wind direction, 9. Wind velocity, 10. Barometric pressure, 11. Pressure change and 12. Clouds and their movement.

Ceiling—The determination of ceiling height is perhaps one of the most important factors of a weather observation for airline purposes. The equipment of the observing station should include a ceiling light for determining the ceiling at night, and ceiling balloon equipment for determining it by day.

Temperature—Accurate thermometers for determining the temperature of the air, and hygrometers for determining the dew point are necessities at all airway stations.

Wind—The wind direction and velocity at the surface are important both for pilots who are to make landings and for meteorologists in analyzing weather charts. An accurate wind vane and anemometer, both arranged for indicating at a distance are essential.

Barometric Pressure—Accurate measurements of the atmospheric pressure are very necessary both for pilots and for meteorologists. It is desirable that both a mercurial barometer and an accurate barograph be available at all airway stations in order to insure accurate pressure observations. If possible the barograph should be of an open-scale type so that pressure tendency observations may be reliable.

Weather observation stations which do not include instrumental equipment indicated above are greatly handicapped and weather reports made by them are subject to considerable doubt.

Supplemental Equipment—Airway stations whose reports are also used for other purposes should include among their equipment maximum and minimum thermometers and a rain gauge. Other equipment which is useful but not essential, includes a nephoscope for cloud measurements and registering devices to provide permanent records of wind direction and velocity, temperature, sunshine, etc.

Upper Wind Observations—To provide information on the direction and velocity of upper winds it is desirable that as many stations as possible be equipped with pilot balloon equipment. In general such stations should be located not farther apart than 150 to 200 miles in order to provide a complete picture of the upper air velocity field.

Aerological Data—Facilities should also be provided at selected stations for the obtaining of upper air aerological information either by regular daily airplane flights or by the use of radio meteorographs so that sufficient information may be available at forecasting centers for the complete analysis of the daily weather charts. These stations should be located not farther apart than 300 to 500 miles.

Dissemination of Data—Since rapid dissemination of weather information is of the utmost importance to airline operation and to the aeronautical forecaster, some of the features of the transmission system employed by the Civil Aeronautics Board in the United States will be described. The service maintained by this agency

is in many respects a model of reliability, thoroughness and efficiency. The transmission facilities of the Civil Aeronautics Board are used to disseminate the information obtained by weather observers, most of whom are members of the United States Weather Bureau. Hourly weather observations are taken at a large number of selected points along the airways within the United States, each of which observation points is in continuous communication with other observation stations along the airway on which it is located. The inter-communication between the various stations along each airway is maintained either by radio or by teletypewriter service.

An example of the arrangements for the transmission of data along a certain airway will give the most satisfactory picture of the manner in which this system operates. For this purpose the airway from Chicago, Illinois, to Newark, New Jersey, will be taken since it is representative of most other airways in the United States. Along this airway, which is approximately 700 miles in length, are located twenty-one stations at which regular hourly weather observations are made. These observations are made at the same time, at thirty-six minutes past each hour, and include the following elements: 1. Ceiling, 2. State of the sky, 3. Visibility, 4. Obstruction to visibility, if any, 5. Precipitation, if any, 6. Temperature, 7. Dew point, 8. Wind direction, 9. Wind velocity, 10. Barometric pressure and 11. Remarks, which may include information on any element of the weather, of importance to aviation, not included in the above.

Each of the twenty-one stations on this airway are interconnected by a long-line teletype circuit and at each station is located a teletype printer capable of both receiving and transmitting. Thus, if any one of the stations on the teletype circuit writes a weather report on its teletypewriter this information will appear on all other teletype machines in the circuit. The circuit is continuously available throughout the day and night for the exclusive use by the stations located on the airway.

Promptly at forty-one minutes past the hour a "sequence heading" is placed upon this teletype circuit by a designated station on the airway. This heading indicates the airway to which it applies and the time at which the sequence starts. Thus, in this case the heading will read, "CG NK 1441CS," indicating that the sequence to follow will cover the Chicago-Newark airway and that the weather observations are taken just prior to 2:41 P.M., 90th meridian time. The time is given according to a twenty-four hour clock with the

time zone also being indicated. (ES—Eastern Standard, CS—Central Standard, MS—Mountain Standard, PS—Pacific Standard).

Immediately following the sequence heading, the first station on the airway will place its weather report on the teletype circuit using a special code developed for this purpose to save as much time as possible. (This code is completely described in "Circular N" of the Aerological Division of the United States Weather Bureau.) As soon as the first station on the teletype circuit has completed transmitting its weather report, the second station places its report on the teletype circuit, then the third station and so on, until the entire twenty-one reports have been placed on the teletype circuit. This complete process generally consumes only two to four minutes. In this manner a complete picture of the weather conditions along an entire airway is obtained in an extremely short space of time. In order to make it possible for weather information obtained on one airway to be transmitted for use by meteorologists and pilots located along other airways, the information obtained may be relayed by central relay stations to other airways. By a complicated series of relays it is thus possible to interchange information obtained on various airways so that complete hourly weather information for practically half of the United States is available on every airway.

In some regions inter-communication between airway stations is maintained by radio instead of teletype circuits. This method of communication is successful during most of the year, but it is subject to occasional interruptions due to static conditions during the summer. The same procedure is followed on these radio circuits as on the teletype circuits. Each station on a certain radio circuit thus obtains the weather for all other stations on the circuit and by the use of a special code it is possible to attain approximately the same speed in transmission as is reached in teletype circuits. In radio airway circuits all stations on a given airway operate on the same frequency. Different airway circuits are established on different frequencies, so it is possible for weather reports to be transmitted simultaneously over several networks of stations.

Since airway reports do not cover the entire area of the United States, additional reports are obtained from stations not located on regular airways. These reports are telegraphed to central distributing points from which they are transmitted over the regular teletype circuits throughout the country. Such off-airway reports are generally obtained at only 6-hourly intervals. On all 6-hourly

weather reports, certain additional information relating to cloud types and pressure change characteristics are added to the regular hourly reports, for use by the forecaster.

In order to complete the weather information necessary for the construction of synoptic weather charts, weather reports are also received by radio from ships at sea in both the Atlantic and Pacific oceans. These are collected by the United States Navy and are placed on the national teletype circuit by centrally located relay stations of the Civil Aeronautics Board. Additional weather reports are also received from Canada and from Mexico by telegraph and by radio and these are also placed on the national teletype circuit by certain designated relay stations.

All the various types of weather reports mentioned above are now coded in one of two codes. Hourly weather reports along the U. S. Airways are coded in the special code described in "Circular N" of the U. S. Weather Bureau. All other surface weather reports are coded in the International Code as adopted by the U. S. Weather Bureau. This code is fully described in a publication of the U. S. Weather Bureau entitled "Weather Code (Numeral System)." This code is slightly different from the code used by ships at sea because of the slightly different material reported in the two cases. The ship code is also obtainable from the U. S. Weather Bureau.

With the present highly developed communications system of the United States Weather Bureau and the Civil Aeronautics Board the meteorologist at any major forecasting center within the United States has available between 150 and 300 weather reports for the construction of each weather map. This widespread and complete weather service is unsurpassed in the world. In some portions of Europe a denser network of stations exists but the area covered and the frequency of the weather reports is not comparable to those obtained in the United States. It is believed that teletype inter-communication is the only entirely satisfactory means of rapidly disseminating weather reports for airline purposes. All other forms of communication including radio telephone and radio telegraph, as well as the ordinary telephone and telegraph, are either subject to errors or are incapable of attaining sufficient speed to allow the establishment of networks of twenty to thirty stations along a single airway, as is necessary to provide complete weather information.

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CHAPTER 3

THERMODYNAMICS

INTRODUCTION

The atmosphere may be considered as a great heat engine which utilizes the natural reservoirs of heat and water provided by the sun and the oceans to produce practically all meteorological processes. Because of the great importance of heat in meteorology it is vital that the serious student of meteorology be thoroughly familiar with at least the basic principles of thermodynamics. This knowledge simplifies the understanding of many natural phenomena otherwise difficult to explain and serves as a useful tool for attacking both the problems of the practical forecaster and those of the purely theoretical meteorologist.

TEMPERATURE

Adiabatic and Isothermal Changes—Transformations involving heat and temperature are most conveniently studied when they are either "isothermal," in which the *temperature* remains constant although the heat content may change, or "adiabatic," in which the *heat content* remains constant but the temperature varies. Many atmospheric processes closely approximate the adiabatic, since the portions of the atmosphere that are being considered are generally large enough so that changes within them are but little affected by their surroundings. Of course, practically no atmospheric process is strictly adiabatic. Ordinarily some heat is gained or lost by radiation or through turbulence effects during the temperature changes. For practical purposes, however, it is satisfactory to regard most changes which occur during the lifting or sinking of large masses of air as adiabatic processes.

Gas Laws—Since the atmosphere is a mixture of gases, it follows the well known laws of Charles and Boyle.

1. The volume of a gas is directly proportional to its absolute temperature, $V = cT$
2. The pressure exerted by a gas is inversely proportional to its volume, $P = \frac{c}{V}$

Combining these two laws, the general gas equation is obtained:

$$PV = RT, \tag{1}$$

Where P is the pressure, V the volume, T the absolute temperature, and R a "gas constant" whose value depends on the units employed. (See Table 8 for values.)

This relation holds strictly true only for so-called, *perfect gases*. Actually some deviation from it occurs with real gases at average pressures and temperatures, but the practical importance of this deviation is small enough to be neglected in most meteorological considerations. When dealing with the same gas and using the same units, the equation may be written, simply,

$$\frac{P_1V_1}{T_1} = \frac{P_2V_2}{T_2} \tag{2}$$

and this relationship holds for all changes involving the three elements of pressure, volume and temperature.

Potential Temperature—For meteorological purposes it is often desirable to compare the temperatures of the air at localities which are at different elevations, and thus at different pressures. Since, as has been shown above, the temperature of the air is directly proportional to its pressure, it is obvious that it will be greatly affected by changes in elevation. If the temperatures of the air at various elevations are to be compared, therefore, it is essential that all temperature readings be reduced to a common pressure. If this be done adiabatically (without gain or loss of heat), the resulting temperature, when the 1000 millibar level is used as the common pressure, is termed the *potential temperature* (θ). This may be calculated readily by the use of the following equations,

$$\theta = \frac{T}{\left(\frac{p_1}{P}\right)^k} \tag{3}$$

$$k = \frac{\gamma - 1}{\gamma} = 0.288 \tag{4}$$

where T is the absolute temperature, p_1 the pressure in millibars, P the standard pressure (1000 millibars), and k a constant involving

γ, the ratio of the specific heat of air at constant pressure ($c_p=0.2396$) to the specific heat at constant volume ($c_v=0.1707$). This value of k should be corrected for the presence of water vapor, although this is of nearly negligible importance.

Within the lower levels of the atmosphere, the adiabatic change in temperature amounts to about 5½° F. per 1000 feet. (1° C. per 100 meters.) This means that if no heat is gained or lost, a particle of air will increase in temperature by about 5½° F. if it is lowered 1000 feet. Conversely it will cool off by the same amount if it is raised 1000 feet, *unless condensation occurs.* After condensation occurs, the rate of cooling becomes much less since the heat furnished by the condensation process tends to warm up the air as it rises.

The rate of cooling of rising, saturated air varies with the water vapor content. With very dry air the rate of cooling, after saturation occurs, is very nearly that of unsaturated air, 5½° F. per 1000 feet. With very moist air, on the other hand, such as might be found on a hot tropical afternoon, the rate of cooling for the saturated state is but 2¼° F. per 1000 feet.

Partial Potential Temperature of the Dry Air—In order to avoid the correction for the presence of water vapor in the equation for the potential temperature, a slightly different quantity is more commonly employed. This is the *partial potential temperature of the dry air,* or simply θ_d. This is obtained from an almost identical equation,

$$\theta_d = \frac{T}{\left(\frac{p_d}{P}\right)^k} \tag{5}$$

where p_d is the partial pressure of the *dry air,* obtained by subtracting the vapor pressure from the total pressure, and where the other quantities have the same meaning, as in equation 3. Table 14 gives the values of the correction to be applied to the ordinary potential temperature, θ, to give θ_d.

The interesting and useful thing about both θ and θ_d that makes them especially useful to the meteorologist, is that they are not at all affected by changes of pressure as long as no condensation occurs. Thus, a particle of air which starts at sea level with a certain temperature will be cooled as it is lifted over the land at the rate of 5½° F. per 1000 feet. Using *air temperature* alone, it would

obviously be very difficult to recognize the particle after it had ascended and become cooler. If, however, the *potential temperature,* (θ), is used in reporting the temperature of the particle, it would make no difference whether the particle went up or down—it would always retain the same potential temperature as long as no condensation occurred. The same would hold true for the partial potential temperature of the dry air, (θ_d).

Conservative Properties—A property of the air, like the potential temperature, which is not affected as the air moves up and down over the earth, is called a *conservative property.* Clearly, it is greatly different from a non-conservative property, like the temperature. To the meteorologist it is indispensable for it enables him to compare bodies of air at widely different elevations. He can see, by the use of conservative properties, that air over the plateau regions of the western United States is identical with air in the Mississippi Valley. Their *temperatures* may differ by 30° F., but their *potential temperatures* will be the same.

Other conservative properties will be described later. Some of them are useful in many ways, others are employed only in constructing technical diagrams. They are all useful to the student of air masses, though, for he is primarily concerned in studying the behavior of portions of the atmosphere as they move about over the earth.

Equivalent Potential Temperature—It has been mentioned above that the potential temperature (θ_d) remained unchanged during vertical movements of the air as long as no condensation occurred. There is still another expression for the temperature, however, which remains unaltered *even during condensation.* This is the *equivalent potential temperature* or θ_e. θ_e is obtained by first lifting a particle of air to infinity (zero pressure), thus removing all of its moisture, and then lowering it to sea level (1000 millibars). With this unit of temperature it makes no difference whether some of the moisture in a particle of air is removed by condensation, or whether moisture from the outside is added by evaporation—the particle is still recognizable and will still have the same value of θ_e.

θ_e may be determined by the use of the following equation,

$$\theta_e = \theta_d e^{\frac{Lw}{c_p T}}$$

where e is the base of the Naperian logarithms, L the latent heat of condensation, w the water vapor content in grams of water per gram

of dry air, and c_p the specific heat of dry air at constant pressure. This formidable calculation may be obviated by the use of table 15, which readily transforms θ_d to θ_e. θ_e may also be obtained very readily by the use of a Rossby diagram if both θ_d and W are known.

The various expressions for temperature may be illustrated by the use of figure 19. Here the wind is blowing across the mountain range from west to east. At the shore line a body of fresh maritime air arrives with a surface temperature and dew point of 59° F. and 48° F. respectively. This air is lifted as it blows up the mountain and the air cools at the dry adiabatic rate of 5½° F. per 1000 feet. It will reach the point *B* at an elevation of 1000 feet with a

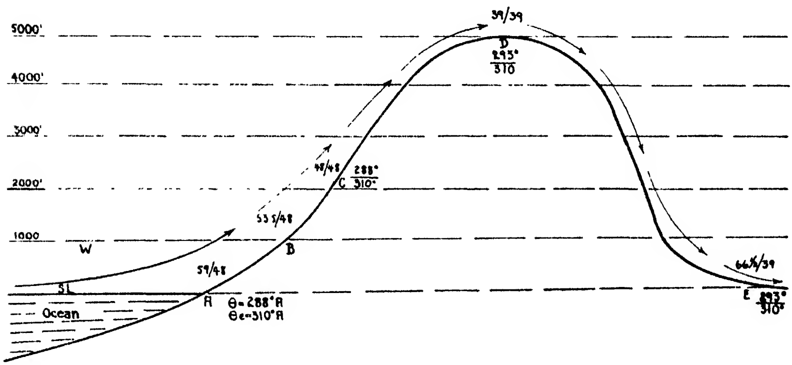

FIGURE 19.—CHANGE IN VARIOUS PROPERTIES OF A MASS OF AIR MOVING ACROSS A MOUNTAINOUS REGION

temperature of 53½° F. and the point *C* with a temperature of 48° F.

The dew point will remain essentially unchanged so far, at 48° F. The potential temperature (θ), and the partial potential temperature of the dry air (θ_d), will not change at all, remaining at the values they had at *A*. Above *C*, the air is saturated and will cool at the *moist adiabatic* rate, in this case about 3° F. per 1000 feet. At the summit of the mountain, *D*, the air will thus have a temperature 9° F. cooler than at *C*, or about 39° F.

If the potential temperature of the air at the summit were now calculated it would *not* be the same as the θ originally at *A*, since some condensation has occurred. Actually the potential temperature of this air would be considerably *higher* than the original

air. Thus the θ for the original air would be 288° A, while for the air at the summit of the mountain it would be 293° A. The *equivalent potential temperature,* (θ_e), however, remains absolutely unchanged during the lifting and condensation processes, 310° A.

As the air now descends the lee slope of the mountain it will be compressed and its temperature will rise at the dry adiabatic rate, 5½° F. per 1000 feet. When it reaches *E* its temperature will thus be 66½° F., considerably warmer than when it started its ascent at *A*. This considerably higher temperature is due to the heat added to the air during the condensation process. θ will be unchanged at the leeward base of the mountain from its value at the summit, 293° A. θ_e is also unchanged at 310° A. θ_d is almost the same as θ throughout, being about 1° warmer.

In summary:

1. The temperature changes with every change in elevation. For the dry state, the rate of change is 5½° F., per 1000 feet. For the saturated state, the rate of change varies from nearly 5½° F. per 1000 feet for very dry air to only 2¼° F. for very moist air.
2. The potential temperature, (θ), and the partial potential temperature, (θ_d), remain unchanged as long as no condensation occurs. After condensation, they increase in value.
3. The equivalent potential temperature, (θ_e), remains unchanged even during condensation.

Wet-Bulb Potential Temperature—Another conservative property which may be used in discussing air masses is the *wet-bulb potential temperature.* This may be obtained readily by the use of an ordinary adiabatic chart which includes the saturation adiabats. The chart is entered using, (1) the wet-bulb temperature from a psychrometer reading and, (2) the station pressure. This point is then carried down the saturation adiabat to standard pressure (1000 millibars). A wet-bulb temperature of 10° C. and a station pressure of 915 millibars, yields a wet-bulb potential temperature of 13.5° C. or 286.5° A. (See figure 20.)

The wet-bulb potential temperature remains unaltered during changes in pressure and also during *both evaporation and condensation.* It may be used similarly to the equivalent potential temperature since it is conservative to the same degree. The changes

in either the wet-bulb potential temperature or the equivalent potential temperature due to heat added from the outside are generally more important by far than those due to evaporation, however, so the latter effect may be neglected in most instances. Since the equivalent potential temperature is somewhat more readily obtained from aerological soundings than the wet-bulb potential temperature it is generally used in air mass discussions. The wet-bulb potential temperature has had but little acceptance by meteorologists generally.

HUMIDITY

The moisture content of the air is of the greatest interest to all those who are concerned with the weather. It determines the possibility of rain or snow, it frequently decides whether an air mass is to be stable or unstable, it profoundly affects the climate of every point on the earth. Because of this, most diagrams used in studying air masses include the humidity in one form or another, and all air mass descriptions mention it.

There are several ways of expressing the humidity or moisture content of the air. One of these is useful because it may be obtained directly from indicating devices, another lends itself readily to various calculations, still another is unaffected as the air is heated or cooled. Each one of these will be described below.

Vapor pressure is at the base of practically all humidity calculations. According to Dalton's Law,

> *The pressure exerted by a mixture of gases is equal to the sum of the separate pressures which each gas would exert if it alone occupied the whole volume.*
>
> $PV = V\ (p_1 + p_2 + p_3 \ldots)$.

Thus the pressure exerted by the water vapor in the atmosphere is a distinct part of the total atmospheric pressure. The vapor pressure may be measured in inches of mercury, millimeters of mercury, or millibars, just as the atmospheric pressure.

For each temperature there is a certain vapor pressure, called the *saturation vapor pressure*, e_m, which is the highest value it may attain without reaching super-saturation. At the freezing point, the saturation vapor pressure amounts to 0.1803 inch or 4.58 millimeters, or 6.11 millibars. At 70° F. it equals 0.7399 inch, or

over four times the value at 32° F. If the atmospheric pressure is 30.0000 inches at a time when the air is saturated at a temperature of 70° F., the *partial pressure of the water vapor* will be 0.7399 inch and the *partial pressure of the dry air*

$$29.2601 \text{ inches } (30.0000 - 0.7399).$$

Relative humidity presents the moisture content as a percentage, in terms of the vapor pressure,

$$\text{R.H.} = \frac{e}{e_m} \times 100$$

where R.H. is the relative humidity, e the vapor pressure, and e_m the vapor pressure at saturation.

When the air is completely saturated, e will equal e_m, and the relative humidity will be 100%; when the air is perfectly dry, e will equal zero, and the relative humidity will be 0%. Intermediate values of the vapor pressure between zero and the saturation point will thus be given as percentages of the latter.

Relative humidity is widely used in many practical ways. In air-conditioning, for instance, the relative humidity is very important in determining a healthful value of the moisture content of the air. In general, it should lie between 40% and 70%. During the winter months the relative humidity of the outside air in the middle latitudes is usually fairly high, ranging from 60%–80%. When this air is heated, however, the relative humidity falls very rapidly. In a house heated to 75° F., the relative humidity may thus be less than 10%, while it may be 70% outdoors where the air temperature is near 0° F. This example emphasizes the desirability of air-conditioning during the winter months to promote healthy conditions of indoor humidity, and also shows the great fluctuations of relative humidity with temperature. Since these variations in the relative humidity occur with no change in the air at all, except heating, it is apparent that this measure of the humidity is not at all conservative. It is thus very poorly suited to the study of air masses.

Absolute humidity expresses the moisture content as the *weight* of water contained in a given *volume* of air. Since the volume of air is inversely proportional to its pressure, this unit of humidity is also non-conservative and practically valueless for air mass studies.

The absolute humidity is not directly obtainable from hygrographs as is the relative humidity, and it has very little practical application.

Specific humidity is a measure of the *weight* of water in a given *weight* of air. It is very useful to the meteorologist since it is unaffected by changes in the temperature or pressure of the air (as long as no water is added or subtracted by evaporation or condensation). The specific humidity may readily be calculated if the relative humidity, the temperature, and the pressure are known, using the following formula,

$$q = 0.622 \cdot \frac{e_m}{p} \cdot \text{R.H.}$$

where q is the specific humidity in grams of water per gram of air, e_m the saturation value of the vapor pressure at the existing air temperature, p the atmospheric pressure, and R.H. the relative humidity.

In several ways, another expression for the humidity is useful. This is the *mixing ratio*, obtained from,

$$w = 0.622 \times \frac{e_m}{p_d} \times \text{R.H.}$$

where w is the mixing ratio in grams per gram of *dry* air, and p_d is the partial pressure of the *dry* air. This is almost identical to q as will be seen from an inspection of the two equations. The only difference lies in the fact that q refers to the moist air, while w refers to the dry air. For all practical purposes the two expressions may be used interchangeably, for the difference, even at high humidities, never exceeds 2 or 3 per cent. This is less than the probable error of most humidity readings.

Generally the specific humidity is given in grams of water per *kilogram* of dry air, (W). (The term specific humidity will be used throughout this book interchangeably for q and w.) Common values for W at the surface are 20 to 25 for very moist tropical air, 7 to 12 for polar maritime air in the summer, 3 to 7 for this type in the winter, and 1 to 3 for polar continental air in the winter. Further details concerning air mass properties may be found in chapters 5, 6, 7, 8.

Dew Point—A useful expression for the humidity, particularly when dealing with fog possibilities, is the *dew point*. This is the

temperature at which condensation will occur if the air is cooled, without change in pressure. It varies with the moisture content, but not with the temperature. It is only slightly affected by changes in pressure. It is thus a fairly conservative property. Because it is little affected by changes of elevation of several thousand feet, it is very useful in air mass identification if the specific humidity is not readily available.

TABLE 2

VARIATION OF DEW POINT WITH PRESSURE

W \ dz	100 mb	200 mb	300 mb	400 mb	500 mb
1	0.9	0.9	0.8	0.8	0.8
5	1.0	1.0	1.0	0.9	0.9
10	1.2	1.1	1.1	1.1	1.0
15	1.3	1.2	1.2	1.1	1.1

For given values of specific humidity *W*, and total vertical displacements *dz*, the tabulated values give the rate of decrease of the dew point in °F. per 1000 feet increase in elevation. The starting point for the lifting process is the 1000 millibar level. The error involved in starting the lift at other elevations is inconsequential, however.

Table 2 gives the average rate of change of the dew point for various values of the specific humidity, and the total vertical displacement, commencing at the surface. It will be seen that for displacements of 1000–3000 feet the change of the dew point amounts to less than 2° F., for moderate values of *W*. For all practical purposes, therefore, the dew point may be considered as essentially conservative for air mass comparisons at the surface. In this connection it should be noted that the above table may *not* be used in comparing dew points at adjacent mountain and valley stations. In the case of air flowing smoothly over a gently rising surface, however, the table shows that the dew point will change very little with changes in pressure of the order of two or three inches. Actually the changes in the dew point involved in a vertical displacement of 2000–3000 feet are well within the limits of error of all but the most accurate humidity observations.

THERMODYNAMICAL DIAGRAMS

In order to study the structure and behavior of the atmosphere it is necessary to use diagrams to show the distribution of the various atmospheric elements. Many such diagrams have been used from time to time, each one of them having its special advantages. Possibly the simplest diagram is one in which temperature and altitude are plotted on linear rectilinear coordinates. The elevation of inversions and other information regarding the temperature lapse rate may be determined readily from this simple chart (figure 104).

Adiabatic Chart—An improvement on the elementary temperature-altitude chart is the *adiabatic* chart. On this, the ordinates are the 0.288 powers of the pressure decreasing upwards and the abscissae are the temperatures on a linear scale, increasing to the right. Sloping lines of constant *potential* temperature (*dry adiabats*) are also drawn. The processes are assumed to be adiabatic, with no gain or loss of heat from the outside.

With this chart it is possible to tell at a glance whether a given portion of the atmosphere is stable or unstable in its *unsaturated* state. Nothing can be inferred about the saturated state unless lines are drawn to give the temperature-pressure relations for saturated air (*saturated adiabats*). Figure 20 represents an adiabatic chart with both dry and saturated adiabats.

The adiabatic chart is very useful in studying the behavior of a particle of air which is displaced vertically. As long as it remains unsaturated it will follow the dry adiabat through its initial position. Thus a particle of air at A (figure 20) will move along the dry adiabat BC if it is raised or lowered in the atmosphere. If it is raised to the 700 mb level its temperature will fall from T_2 to T_1, as it moves from A to A', as long as it remains unsaturated.

As soon as the particle becomes saturated, however, the heat of condensation will retard the adiabatic cooling and the particle will follow a *saturated* adiabat. If the particle moving upward along BC becomes saturated at A' it will then proceed up along the *saturated adiabat* through A'. It will then reach A'' at the 600 mb level. Here it will have a temperature T.

If the particle descends it will always follow the dry adiabat. Thus, the particle at A'' will return along the dry adiabat to A''' if it descends to its initial position. If it becomes saturated during its ascent it will return down along a different dry adiabat than it

followed in its ascent. The particle at A'' will thus follow the dry adiabat DE in its descent.

It is apparent from a consideration of the adiabatic chart that the temperature of a particle which is lifted until condensation occurs

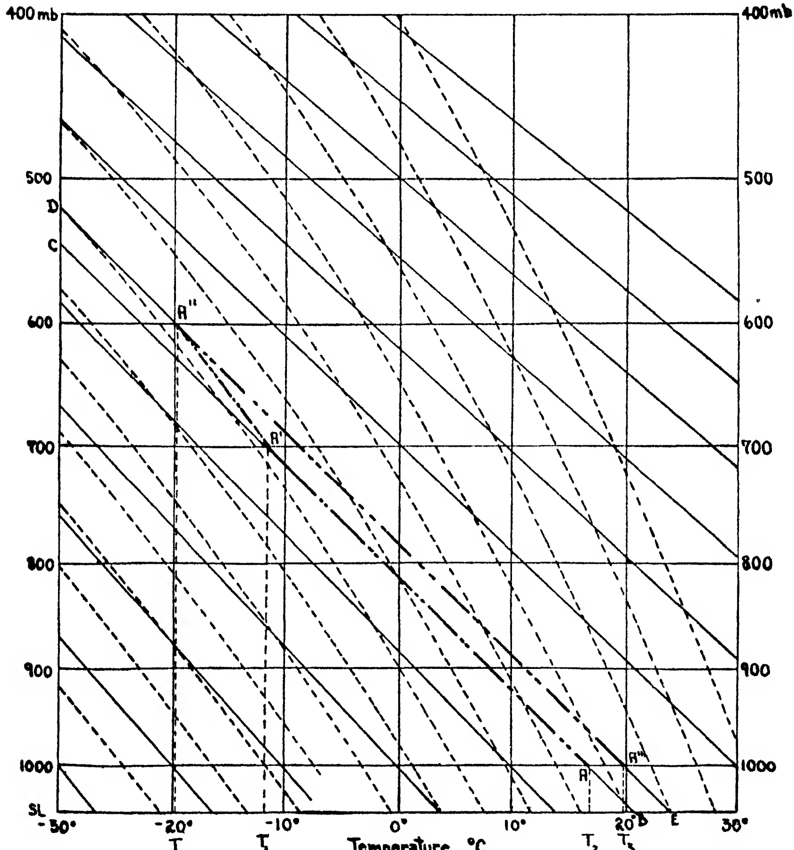

FIGURE 20.—ADIABATIC CHART

Vertical lines represent constant temperature in °C.; horizontal lines represent constan pressure in millibars; sloping solid lines are dry adiabats; sloping dashed lines are saturated adiabats.

(from A to A'), and then further lifted after condensation (from A' to A''), will be warmer than it was originally, when lowered to its original level (from A'' to A'''). The temperature increase will be the difference between T_3 and T_2.

Stability—The *stability* of an air mass is the resistance that it offers to vertical movement of particles of air within it. It may be explained most readily by the use of an adiabatic chart. In figure 21 the curve AB represents the temperature distribution of a certain air mass. If a particle of air at A is lifted it will follow the dry adiabat through A until it is saturated (A to A_1). It then

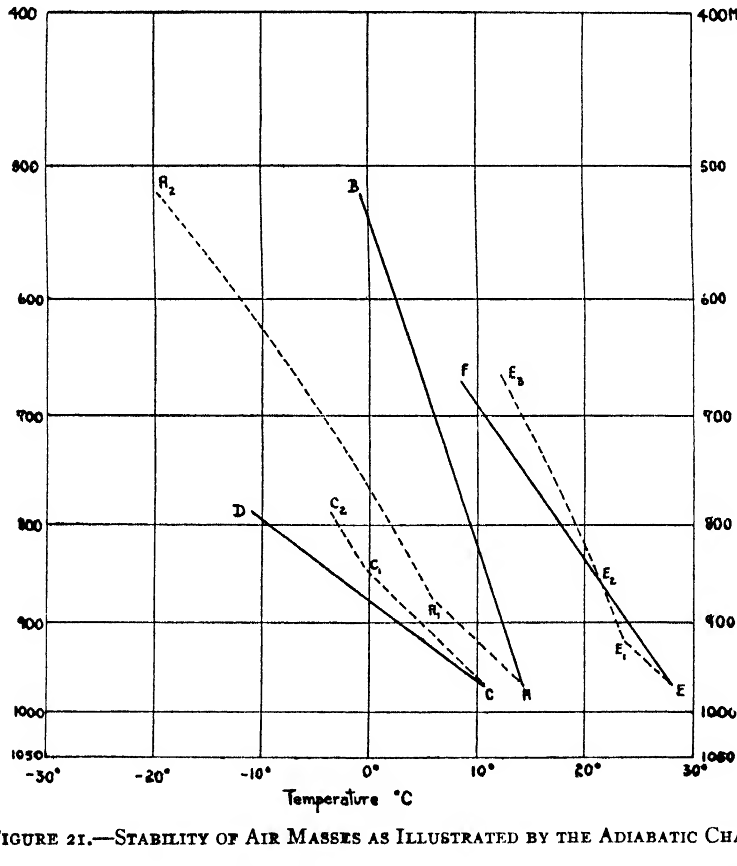

FIGURE 21.—STABILITY OF AIR MASSES AS ILLUSTRATED BY THE ADIABATIC CHART

will follow the saturated adiabat through A_1 (A_1 to A_2). Throughout this entire process, the rising particle is *cooler* than its surroundings (represented by the air mass curve AB). It therefore *resists* the lifting since it is denser than the air through which it is rising. It may be shown similarly that it resists downward displacement. The air represented by the curve AB is thus said to be *stable*.

Consider now a different air mass, represented by the curve *CD*. If a particle within the air mass, as a *C*, be lifted, it will follow the dry adiabat until it is saturated (C to C_1). Then it will follow the saturated adiabat (C_1 to C_2). During the entire lifting process the rising particle will be *warmer* than its surroundings (represented by *CD*). It therefore rises freely of its own accord once an initial upward impetus has been given it. It also sinks freely once an initial downward impetus has been supplied. This air, represented by the curve *CD*, is said to be ***absolutely unstable.***

In a third case the temperature distribution is represented by *EF*. Here, a rising particle at *E* will follow the dry adiabatic (from E to E_1), until saturation occurs, then it will follow the saturated adiabat (E_1, E_2, E_3). While the rising particle is unsaturated it follows the dry adiabat EE_1 and is cooler than its surroundings. It therefore resists lifting. After it becomes saturated it follows the saturated adiabat. From E_1 to E_2 it is cooler than its surroundings, but from E_2 to E_3 it is warmer than its surroundings (represented by *EF*) and rises freely of its own accord. An air mass with a curve similar to *EF*, intermediate in slope between the dry and the saturated adiabats, is said to be ***conditionally unstable.*** This term is derived from the fact that the instability is conditional upon sufficient lift being supplied to saturate the air.

In actual conditions, absolute instability is very uncommon, since air in this condition will tend to rearrange itself and become stable or at least *neutrally stable.* When air is neutrally stable it neither resists nor assists vertical displacements. Generally air is either *stable* or *conditionally unstable.*

Displacements of Entire Layers—The preceding discussion has been concerned only with the lifting or sinking of individual particles. Attention will now be turned to the effects of vertical movements on the stability of air strata of definite thickness. Figure 22 represents an adiabatic diagram with an aerographic sounding plotted. It is proposed to investigate the changes in the stability of various layers of the air mass represented by this sounding when it is raised or lowered as a unit. This type of motion is essentially that which occurs as an air mass moves over a frontal surface. The discussion in regard to lifting applies only so long as no condensation occurs. For a discussion of the behavior of *saturated* air see pages 60-62.

The adiabatic charts used in this discussion have exponential

pressure scales arranged so that equal distances along them represent approximately equal vertical distances. Most air mass movements involve *constant mass transport* rather than constancy of volume. It will therefore be necessary to consider the displacements of layers which possess equal pressure differences at top and bottom, rather than layers of equal thickness.

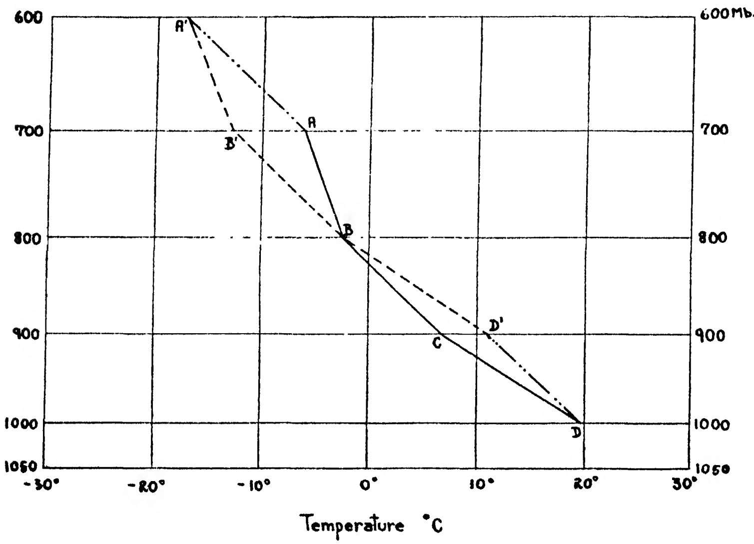

FIGURE 22.—STABILITY OF LAYERS WITHIN THE ATMOSPHERE AS ILLUSTRATED BY THE ADIABATIC CHART

The curve *ABCD* represents a sounding which exhibits various degrees of stability. *AB* represents a layer 100 mb thick which is stable for the dry state (perhaps conditionally unstable, although this is uncertain because the moist adiabats are not shown); *BC* a layer in neutral equilibrium for the dry state (unstable for the saturated state); *CD* a layer that is absolutely unstable for the dry as well as the saturated state.

If the layer *CD* is lifted as a unit a distance of 100 mb it will cool adiabatically (assuming no condensation) until the top and bottom reach the positions shown at *BD′*. It may be seen that the layer remains unstable, but less so than before lifting. (This is shown by the fact that the curve is more nearly parallel to the dry adiabats.) ***An unstable layer becomes less unstable as it is lifted and approaches neutral stability.***

If BC is lifted 100 mb it will reach the position BB', and suffer no change in its stability since the entire layer cools at the dry adiabatic rate.

The layer AB will reach the position $A'B'$ after being lifted 100 mb. It may be seen that this layer remains stable, but becomes somewhat less so than before lifting. (This curve also becomes more nearly parallel to the dry adiabats.) *A stable layer becomes less stable as it is lifted and approaches neutral stability.*

Subsidence—The sinking of large portions of the atmosphere deserves considerable attention because it is a very common phenomenon in nature. The thermal structure of an air mass is often

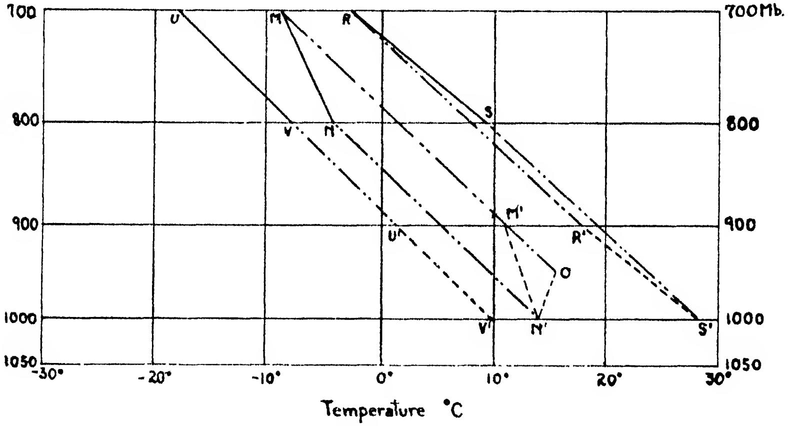

FIGURE 23.—SUBSIDENCE IN THE ATMOSPHERE ILLUSTRATED BY THE ADIABATIC CHART

greatly altered by subsidence, causing a complete change in its stability. Subsidence generally occurs in connection with stagnating high pressure areas and it may involve thousands of square miles. Subsidence is a divergent process, and is generally associated with isallobaric highs and with frontolysis.

The subsidence of a layer of air is best explained by the use of an adiabatic chart, as with lifting processes. In figure 23 a layer of absolutely stable air 100 mb thick is represented by the line MN. If this layer sinks 200 mb, its stability will be increased as is shown by the change in slope of its pT curve ($M'N'$). If, furthermore, it spreads out during the subsidence, so that its thickness decreases to 50 mb, ($N'O$), its stability will be tremendously increased, with

a strong temperature inversion being formed. This is a very common process in nature, and is most important in stabilizing air masses. Most of the inversions which are not caused by radiation effects are due to subsidence. *A stable layer becomes more stable as it subsides.*

Referring again to figure 23, a layer which is initially conditionally unstable is represented by the curve *RS*. If this layer sinks 200 mb, to *R'S'*, it becomes more unstable. *An unstable layer becomes more unstable as it subsides.* Actually this is a rare occurrence in nature, since mixing usually causes the establishment of an adiabatic lapse rate.

A layer in neutral equilibrium will remain unchanged as shown by the curve *UV*. When this sinks to *U'V'* the layer remains in neutral equilibrium, since the entire layer possesses the same potential temperature.

Rossby Diagram—The properties used in constructing an adiabatic chart—pressure and temperature—are not conservative. For this reason, the adiabatic chart is very poorly suited to the comparison of air masses at different stages in their history and at various localities. Professor Rossby has developed a chart which is very well adapted to the study of air masses, since it employs properties that are strictly conservative for the unsaturated state. He uses the partial potential temperature of the dry air (θ_d), and the specific humidity (W) as coordinates. θ_d is represented in absolute temperature on a logarithmic scale and W is given in grams per kilogram on a linear scale (figure 24). Lines of constant equivalent potential temperature (θ_e) are also constructed.

Air mass curves are plotted on the Rossby diagram after calculating θ_d or θ_e, and W. Usually these quantities are calculated at the station making the observations. The tables in the Appendix may be used for making these calculations. W is obtained by using table 12 for the saturated vapor pressure.

$$W = \frac{.622 \cdot e_m \cdot \text{R.H.}}{p_d} \times 1000$$

where e_m is the saturated vapor pressure, R.H. the relative humidity, and p_d the atmospheric pressure. e_m and p_d must be in the same units. W is the specific humidity (mixing ratio) in grams per kilogram of dry air.

θ may be determined from inspection by using an adiabatic chart. θ_d or θ_e may then be found by the use of tables 14 and 15.

On a Rossby chart the instability of any layer of an air mass may be determined by simple inspection. A curve which slopes downward to the left (*AB* in figure 24) represents a layer of air in which the partial potential temperature *decreases* aloft. This layer is thus *absolutely unstable*, since any particle within it, if given an

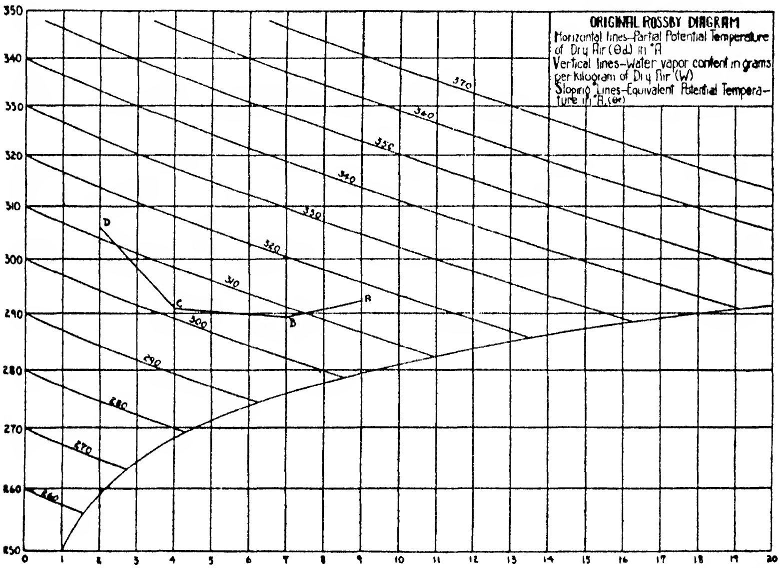

FIGURE 24.—ROSSBY DIAGRAM SHOWING VARIOUS DEGREES OF AIR MASS STABILITY

initial impetus either upward or downward, will continue to move of its own accord.

A curve *BC* represents a layer in which θ_d *increases* with elevation but θ_e *decreases*. This is said to be *convectively unstable*. If the layer is lifted (as by convection), it will eventually become saturated. The layer will then be unstable since the *equivalent* potential temperature decreases with elevation. This state is seen to be very similar to *conditional* instability, discussed in the section on adiabatic charts. Conditional instability involves the behavior of an individual particle, whereas convective instability considers an entire layer.

The situation in which both θ_d and θ_e *increase* with elevation is represented by the curve *CD*. Any particle within this layer will

resist vertical displacements both before and after saturation. This curve therefore represents a *stable layer.*

Absolute instability is very rare, since a layer in this condition tends to readjust itself and become stable. Convective instability and absolute stability are very common and represent practically all natural cases. In the discussion of air masses, the occurrence of various examples of these stability conditions will be described.

It may be noted from an inspection of the original Rossby diagram that the average angle between the θ_d and the θ_e lines is about 20°. Thus only about 1/18 of the total field of the chart is available to represent the slope of the curves corresponding to the very important state of convective instability. Mr. Allan Clark has constructed a slightly different diagram in which the coordinates are, θ_d and θ_e on a linear scale, instead of θ_d and W. This chart has the very important advantage of spreading the field of convective instability over 90°. The revised Rossby diagram also includes lines of constant W.

Figure 44 represents the revised Rossby diagram. θ_d increases from bottom to top and θ_e from left to right. Sloping lines represent W in grams per kilogram. (1) *Characteristic curves which slope down to the left are absolutely unstable* (both θ_e and θ_d decrease). (2) *Curves which slope up to the left are convectively unstable* (θ_d increases but θ_e decreases). (3) *Curves which slope up to the right are absolutely stable* (both θ_e and θ_d increase). Thus, the two stability conditions most frequently encountered each occupy a 90° sector of the chart.

The effects of lifting and sinking on the stability of layers of air may be discussed conveniently with the aid of the revised Rossby diagram. In figure 25, isobars of the partial pressure of the dry air at the condensation level are shown as curves sloping upward to the right. Curves of constant W are not included, since they are not necessary for the present discussion. The characteristic curve of a certain layer is represented by AB. In this layer both θ_d and θ_e increase aloft, showing that it is absolutely stable. If the layer is lifted, its curve will show no change until saturation occurs. After a point in the layer becomes saturated it will move up along a constant θ_e line.

When the layer becomes saturated every point in it will follow constant θ_e lines. If it is then lifted 200 mb, the top and bottom will move to A' and B' respectively. In the new position,

the curve $A'B'$ is more nearly parallel with the θ_e lines. Thus, the layer becomes less stable when lifted *even after saturation.* The limit of this process is neutral stability, with θ_e constant at all levels.

When an absolutely unstable layer is lifted after it has become saturated, its instability increases still further (CD, figure 25). Also, when a layer in neutral equilibrium for the dry state is lifted after becoming saturated it becomes absolutely unstable (EF, figure 25).

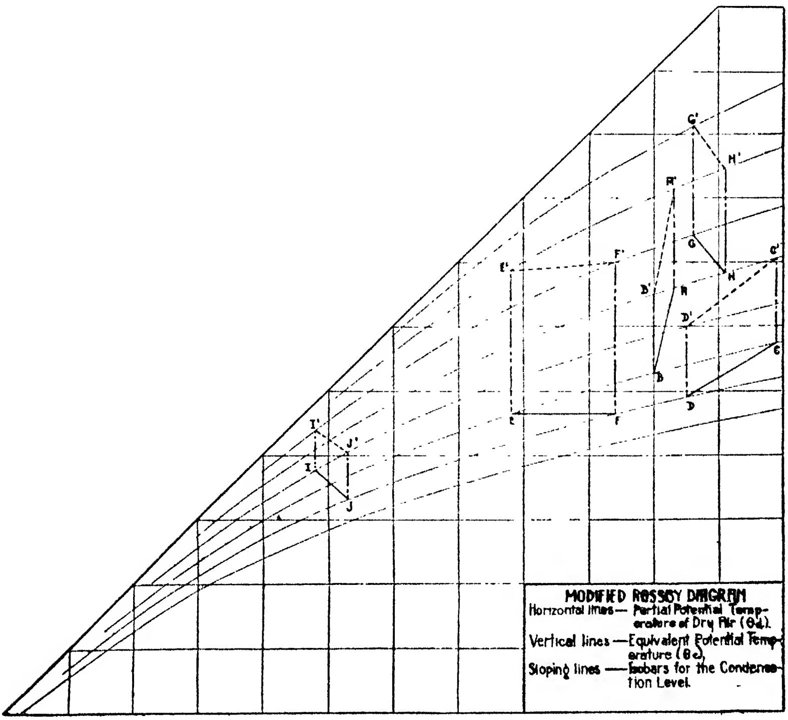

FIGURE 25.—ROSSBY DIAGRAM AS REVISED BY CLARK SHOWING VARIOUS TYPES OF STABILITY CONDITIONS

A layer which is initially convectively unstable generally becomes less unstable when it is lifted. With low values of θ_e and θ_d, however, the layer may become more unstable as it is lifted. Thus GH in figure 25 becomes *less* unstable as it is lifted to $G'H'$, whereas IJ becomes *more* unstable when lifted to $I'J'$.

With the Rossby diagram, *no change whatever occurs in the characteristic curve of a layer of air as it descends*, provided no water or heat is added by turbulent mixing or evaporation, because the

equivalent potential temperature, partial potential temperature, and specific humidity remain unchanged under these circumstances. As has been pointed out above no change occurs when a layer is lifted, until condensation occurs.

The Thetagram—In order to study the day-by-day changes in the atmosphere above a given station, it will be found convenient to plot a diagram showing the distribution of equivalent potential temperature with *elevation*. This type of diagram has been called a *thetagram* by Schinze (figure 26). It is more useful than the

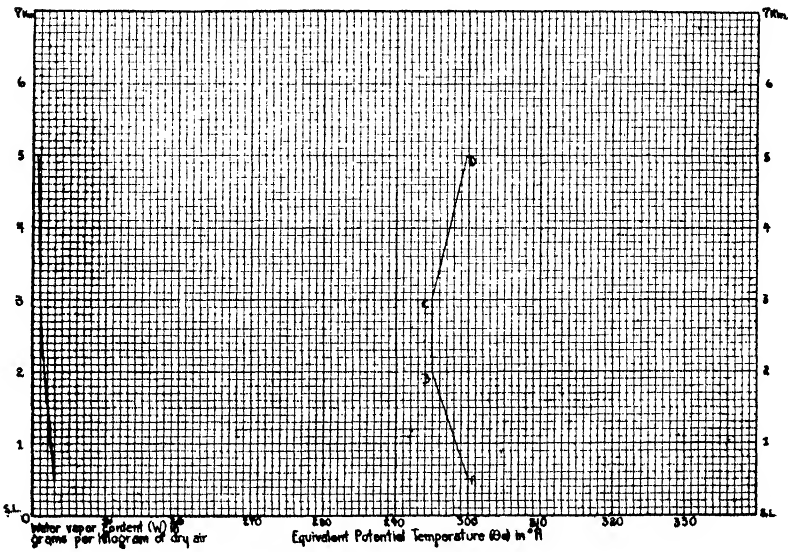

FIGURE 26.—THE THETAGRAM

On this chart equivalent potential temperature is plotted against elevation. Curve or the specific humidity may also be added as shown.

Rossby diagram in studying the *changes* that occur at various levels from day to day at a given locality. It shows directly the *elevations* at which discontinuities are present. It has several marked disadvantages in the study of air masses, since it does not distinguish between the effects of moisture and of temperature distribution (both of these elements are implicit in θ_e). It therefore does not show whether a discontinuity is due to one or the other of these elements. This can be shown, however, by plotting W on a separate diagram to one side, using the same elevation scale.

The thetagram shows at a glance whether any layer is convec-

tively unstable. With the diagram plotted as in figure 26, a slope of the characteristic curve to the left (as at AB) shows that θ_e decreases with elevation and that the layer is in a state of convective

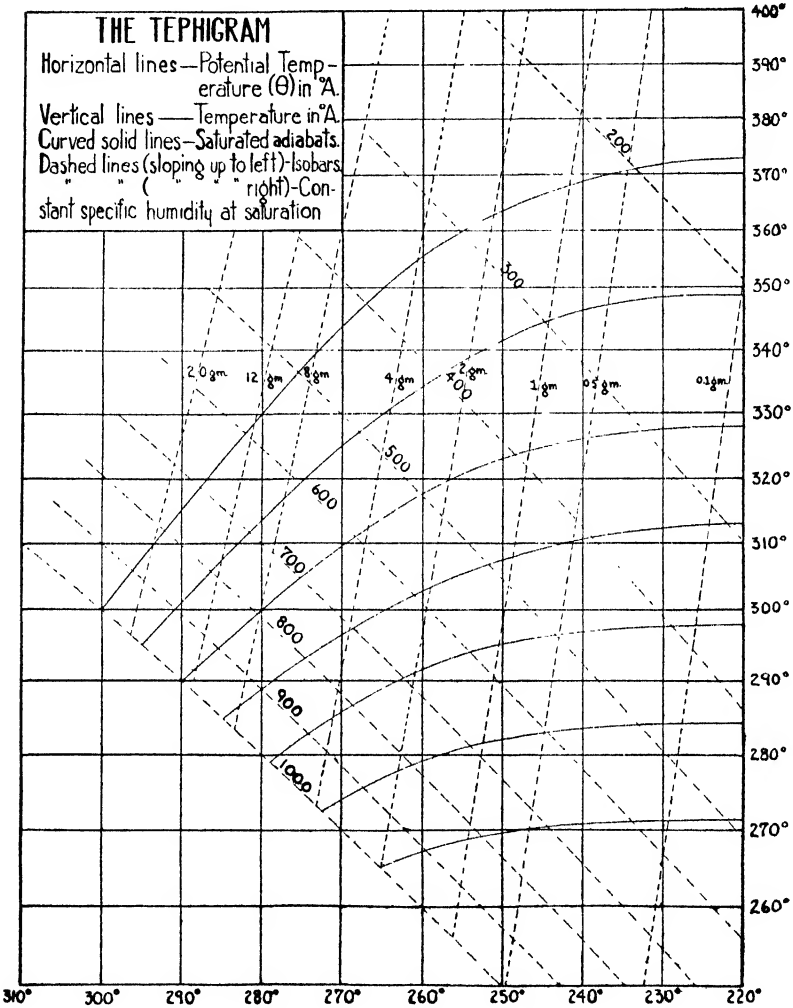

FIGURE 27.—THE TEPHIGRAM

instability. When the curve slopes to the right (as at CD) it indicates that the layer is absolutely stable. When it is vertical, the layer is in neutral stability for the saturated state.

The question of stability is much better treated by other diagrams such as the Rossby diagram or the tephigram, or emagram, but the thetagram retains its value in studying comparative changes in the atmosphere at various elevations, since this feature is not clearly shown by any of the other diagrams.

The Tephigram, figure 27, as developed by Sir Napier-Shaw, is a very satisfactory means of studying by thermodynamical considerations the energy available in the atmosphere for convection. In this diagram, the *temperature* and the *potential temperature* are plotted on rectilinear coordinates, the temperature on a linear scale and the potential temperature on a logarithmic scale. As a result of this method of plotting, equal areas on this diagram represent equal amounts of *work*, since the potential temperature is directly proportional to the entropy. A determination of available energy is thus readily performed.

The Emagram—An improvement on the tephigram, at least for practical purposes, has been made by Refsdal who has developed the *emagram*, figure 28. In this chart the pressure and temperature are plotted as on the usual adiabatic chart, with temperature on a linear and pressure on a logarithmic scale. The dry and saturated adiabats are also indicated, as well as isopleths representing W at the condensation level. As with the tephigram, equal areas represent equal amounts of work.

It is readily possible with this diagram to compute the energy available for convective activity in the atmosphere if an aerographic sounding is available. A temperature-pressure curve is first drawn, as shown by the curve $ABCDEFG$ (figure 86). The energy available for convective activity is obtained by following on the diagram the path of a particle as it is lifted. The particle will follow the dry adiabat through its initial position until it becomes saturated, then it will move up along the saturated adiabat. Referring to figure 86, a particle at point A on being lifted will follow the dry adiabat AH until it becomes saturated at H. The elevation of H is called the *lifting condensation level.* It is located at the intersection of the dry adiabat through the initial point, and the saturation curve having a value equal to the specific humidity of the initial point. After the particle becomes saturated it will follow the saturated adiabat through H.

Since areas on the chart represent *work*, the areas between the characteristic curve of the air mass and the path of the particle

represent energy. This energy is considered *positive* when it assists the displacement of the particle, and *negative* when it retards the movement of the particle. In figure 86, the positive energy is repre-

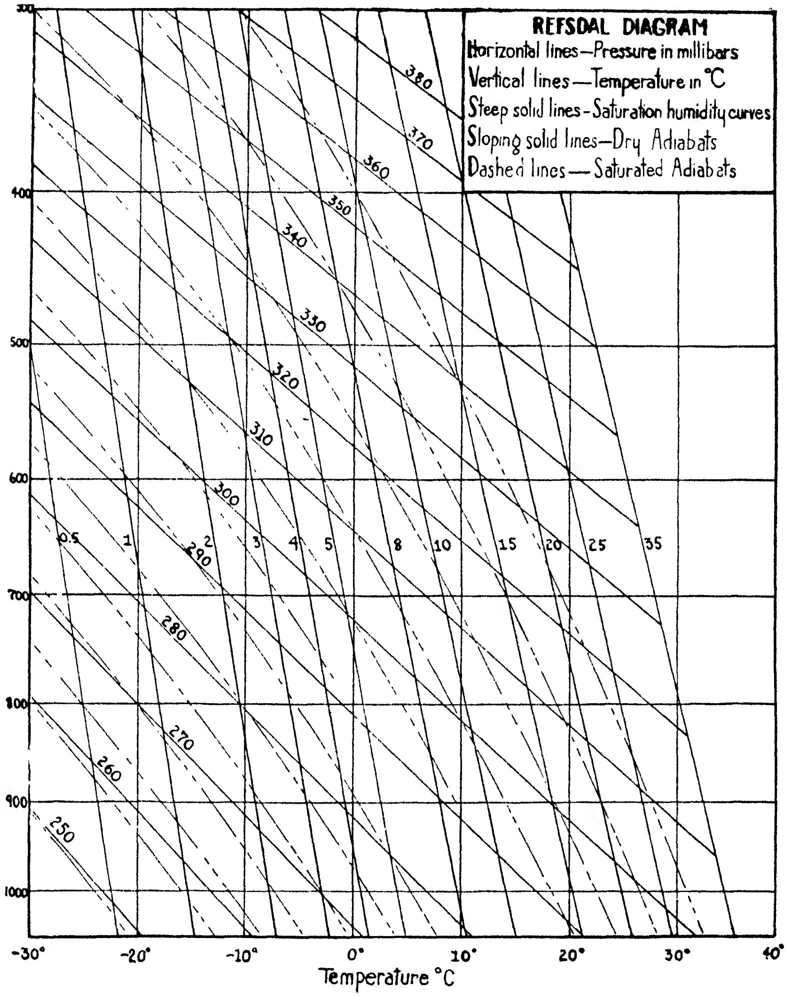

FIGURE 28

sented by the area *KLMFEDK* and the negative energy by the area *KHABCK*. The net energy available for producing instability is the algebraic sum of these two areas. If negative area predominates, there will be little chance for instability to develop. If posi-

tive area predominates, the air mass may become actively unstable.

Another important level is the *convective condensation level.* This must be determined whenever active convection in the lower levels is to be expected. The use of the convective condensation level greatly changes the positive and negative area relations and it is very important to consider when dealing with thunderstorms. This level is indicated in figure 86. It lies at the intersection of the saturation curve corresponding to the average specific humidity of lower levels of the atmosphere and the sounding curve itself. A new emagram may then be constructed using the convective condensation level, instead of the lifting condensation level. It is seen to be greatly different than the emagram which employed the lifting condensation level, although both were constructed from the same sounding.

The use of the convective condensation level is illustrated on pages 238-240 in connection with thunderstorms. In fact the chief use of the emagram is in dealing with thunderstorms, since a knowledge of the energy available for free convection in the atmosphere is indispensable to any approach to this problem. The positive and negative areas give directly the relative amounts of energy available, to aid and to inhibit free convection.

Atmospheric Cross-Sections—Since a knowledge of the vertical distribution of temperature and moisture in the atmosphere is of prime importance in modern forecasting, it is clear that some means is needed to present this information to the forecaster in a rapidly prepared and readily usable form. Willett has succeeded in developing a comparatively rapid means of preparing vertical "cross-sections" of the atmosphere by using the rather complete network of aerological stations in the United States. He draws profiles, both east-west and north-south through convenient aerological stations, and on these he plots at each station the vertical distribution of θ_e and W, as described for the thetagram. Isopleths for these elements may then be drawn, and frontal surfaces will generally be found to be clearly delineated. These cross-sections have the property of showing upper fronts more clearly than they can be shown by ordinary methods, and they also indicate the slope of frontal surfaces. Because of the necessity of exaggerating the vertical scale considerably, the slopes of all surfaces are greatly distorted.

Summary of Thermodynamical Diagrams—The student of

modern meteorology should be thoroughly familiar with all of the diagrams described in this chapter, even though he may not find it practical to use more than one or two in daily forecasting. This is true because even though the practical synoptic meteorologist does not find time to investigate completely all of the thermodynamical properties of the air masses with which he deals in his daily routine, he will form a judgment from a thoroughgoing study of the various diagrams that will enable him to make deductions of great value from a mere cursory perusal of aerographic soundings.

If time and assistance were available the meteorologist would find it desirable to plot daily for the area in which he is interested, the emagram, the Rossby diagram, and the thetagram for each aerographic station, and then to plot several atmospheric cross-sections. The information that these charts would present to him would be invaluable in preparing forecasts, but the labor necessary to prepare them ordinarily makes it necessary to curtail this program except in exceptionally well staffed meteorological offices. In the average meteorological office with one or two meteorologists on duty, it will be necessary to choose certain key stations in the aerological network and to plot thetagrams regularly for them. Emagrams should then be prepared for locations where convective activity may be expected. This is generally all that can be done in the practical daily use of the various thermodynamical charts which have been described. This very fact emphasizes, however, the importance of the meteorologist being thoroughly familiar with all types of diagrams so that he may identify air masses from a glance at their aerographic sounding even before they are plotted on a Rossby diagram. He should be able to tell from a knowledge of characteristic air mass properties where convective activity may be expected.

The use of atmospheric cross-sections is practically impossible for daily forecasting use because of the length of time required for their construction. These cross-sections have their greatest use in training the student to think of the atmosphere in three dimensions, and in becoming accustomed to common configurations of air masses. When considerable ability is obtained by the meteorologist in treating the atmosphere in three dimensions he will generally find that he can picture its structure by means of a study of the surface chart together with isolated aerographic soundings, and that the cross-sections are not absolutely necessary.

REFERENCES

1. Der Feuchtlabile Niederschlag: Anfinn Refsdal. Geof. Publik., v. 5, n. 12, 1929.

This is an excellent treatise on the theory and practical applications of conditional instability as determined from aerographic soundings. A large number of practical examples of the use of these concepts in the forecasting of thunderstorms are presented. The very useful energy diagram, the *emagram* is described. A short section on extra-tropical cyclone structure is also included. This paper is written entirely in German.

2. The Tephigram, Its Theory and Practical Use in Weather Forecasting: Clifford M. Alvord and Robert H. Smith. Mass. Inst. Tech. Meteor. Course, Prof. Notes, n. 1, 1929.

The authors of this paper give in detail the theory and construction of the tephigram. They also present a number of illustrations of its use in forecasting thunderstorms.

3. Thermodynamics Applied to Air Mass Analysis: C.-G. Rossby. Mass. Inst. Tech. Meteor. Papers, v. 1, n. 3, 1932.

The important *Rossby diagram* was first introduced in this paper. The derivation of the more important thermodynamical quantities used in meteorology, such as specific humidity, equivalent potential temperature, etc., is presented here. A few synoptic situations are included to show the usefulness of thermodynamics in practical air mass analysis.

4. Physical and Dynamical Meteorology: David Brunt. New York, The Macmillan Company, 1934.

The sections on thermodynamics of this book are to be recommended.

CHAPTER 4

CIRCULATION

INTRODUCTION

Atmospheric circulation is fundamental to all weather phenomena. On a large scale it carries air from the tropics, laden with heat and moisture, into the middle latitudes where it interacts with cold, dry air from the poles. Most extratropical storms are produced in this way. Storm tracks are determined by the general air movement in various latitudes. On a smaller scale, circulation between land and sea results in fog and mist along the shore. All types of precipitation are the direct results of air movement caused by either general or local circulations. This chapter deals with the causes of circulation, some of the larger features of worldwide circulation, and some of the more interesting types of minor circulations.

THE CIRCULATION PRINCIPLE

It is a well known fact that air tends to flow from a cold region to a warm region. This phenomenon, which accounts for both large and small scale circulations, has been explained in various ways, but perhaps most satisfactorily by V. Bjerknes with the use of *solenoids* and the *circulation principle.* If the field of pressure within a fluid be represented by surfaces of equal pressure (*isobars*), and the field of density by surfaces of equal specific volume (*isosteres*), it is possible to discuss their action on the fluid by graphical means using some of the principles of elementary vector analysis.

In figure 29-a, the isobaric surfaces and the isosteric surfaces of a fluid are parallel. This condition, which is realized more or less approximately by relatively small bodies of liquid, either at rest or in motion, is known as the *barotropic state.* In this illustration the density of the fluid is constant along surfaces which are equidistant from its surface. This is not strictly necessary for the barotropic state, since the only requirement is that the isosteres and isobars be

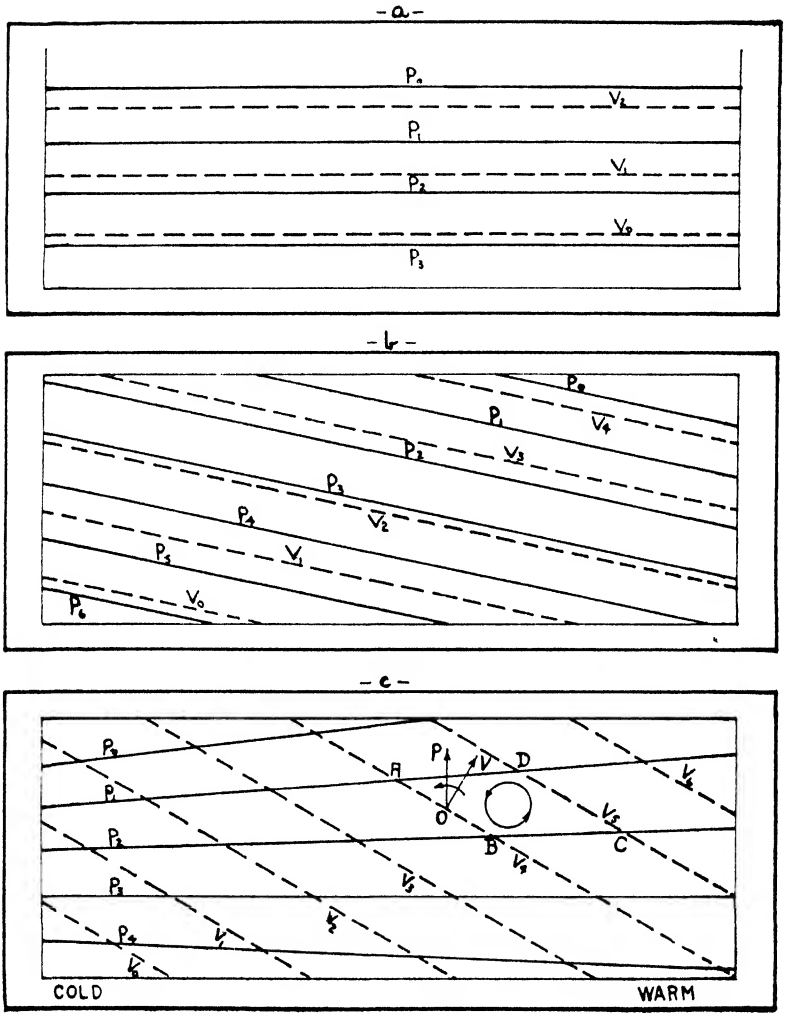

FIGURE 29.—(*a*) BAROTROPIC STATE WITH PARALLEL, HORIZONTAL ISOBARS AND ISOSTERES. (*b*) BAROTROPIC STATE WITH PARALLEL, INCLINED ISOBARS AND ISOSTERES. (*c*) BAROCLINIC STATE WITH INTERSECTING, INCLINED ISOBARS AND ISOSTERES

Note the system of solenoids with circulation occurring as shown in (*c*).

parallel. Both may be inclined to the surface as in figure 29-b. This is a condition that might occur when the fluid is in motion. In a fluid which is in the barotropic state, no tendency exists for circulation between various portions.

If a temperature difference be established between various portions of the fluid in figure 29-c, so that the warmer portion is at the right, the fields of pressure and specific volume will take on the configuration shown. Here the isosteres slope rather steeply to the right showing that the density at a given level is higher toward the left. The isobars diverge slightly to the right since the pressure of the denser fluid to the left is somewhat higher than the lighter fluid to the right, at any given level. The isobars and isosteres are no longer parallel and the fluid is now in the *baroclinic state.* This is the situation usually present in the atmosphere.

The quadrilaterals formed by the intersection of the isosteres and the isobars, as *ABCD,* are called *solenoids.* These may most conveniently be investigated by employing elementary concepts of vector analysis and representing the fields of pressure and specific volume by vectors. *V* and *P* are drawn normal to the isosteres and isobars respectively. They point toward the direction of the greatest *rate of increase* in specific volume, and the greatest *rate of decrease* in pressure, at the point *O* (figure 29-c). For meteorological purposes, the *ascendent* is defined as the maximum rate of *increase* of a function, and the *gradient* as the maximum rate of decrease of a function. *V* thus represents the *specific volume ascendent* and *P* the *pressure gradient.*

In any baroclinic system a circulation will develop which tends to rotate the specific volume ascendent vector into coincidence with the pressure gradient vector. The circulation in figure 29-c, therefore, will take place as shown by the arrows in the solenoid *ABCD.* (See also figure 103, page 293.) The *acceleration* of the air particles in a baroclinic field is directly proportional to the number of unit solenoids included in the field and thus is proportional:

1. to the rate of change of pressure and density, and
2. to the angle between the isosteres and the isobars.

The *velocity* of the air particles will increase until the available accelerating force is balanced by the frictional forces.

Table 18 may be used to obtain an estimate of the accelerating force under various conditions of temperature (specific volume), and

pressure. This gives the number of unit solenoids included in a field of pressure bounded by the 1000 mb level at the base and the p mb level at the top, and in a field of temperature with a horizontal mean temperature difference of T_a-T_b. A solenoidal field extending from the 1000 mb level to the 700 mb level and having a mean temperature difference of 20° C. will thus enclose 2050 unit solenoids (the units are the Meter-Ton-Second).

The equation for the determination of the solenoid number is,

$$N = R(T_a\text{-}T_b) \log \frac{p_o}{p}$$

where N is the solenoid number,

R is the gas constant,

T_a and T_b the temperature in °C. of the warm and cold sources,

p_o is standard pressure (1000 mb),

p is the pressure of the top of the pressure field.

By multiplying the solenoid number by the time in seconds during which the accelerating force acts and dividing it by the length in meters of the curve enclosing the solenoidal field, N will have the dimensions of velocity. It is possible to obtain in this manner a rough estimate of the wind velocity that may be expected under a given solenoidal condition, neglecting frictional forces. If the value of p is 900 mb, T_a-T_b is 10 degrees and the length of the curve enclosing the field is 100,000 meters, then the velocity after 1 hour would be

$$V = \frac{Nt}{L} = \frac{301 \times 3600}{100{,}000}$$
$$= 10.8 \text{ meters per second.}$$

Actually, frictional forces begin to act the moment that a solenoidal field is established, and are very important in controlling the attainable velocity. Any calculations such as the one above must be used therefore only as rough guides to the velocities that may be expected. For times up to one or two hours the results obtained in this manner approach actual values, but for longer times and for extreme solenoidal concentrations, the results obtained are invariably much too high.

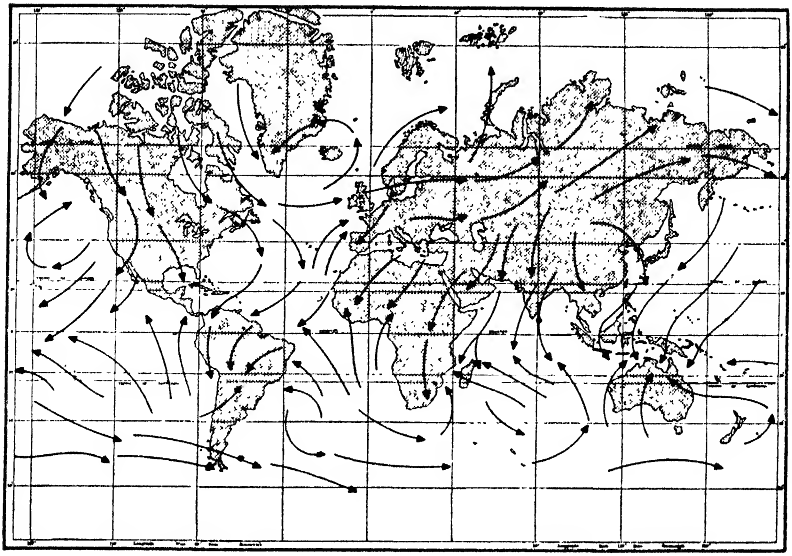

FIGURE 30-*a*. GENERALIZED CIRCULATION OF THE WORLD—JANUARY

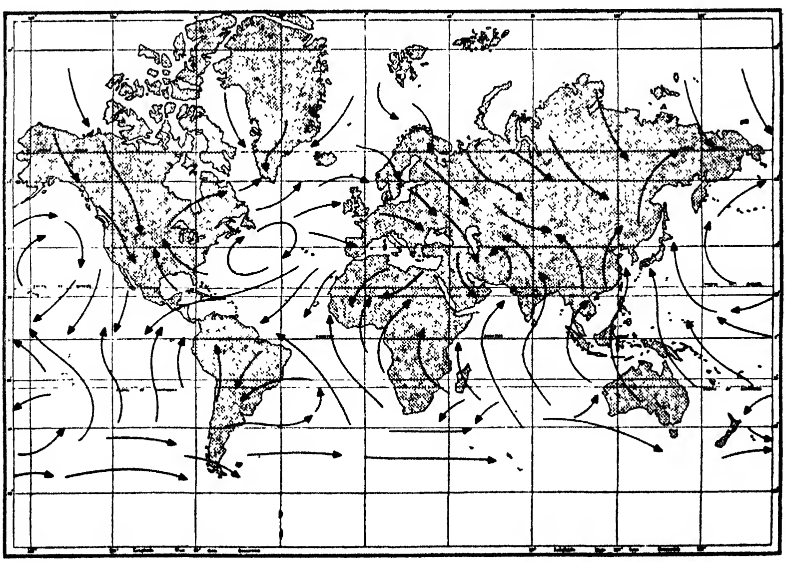

FIGURE 30-*b*.—GENERALIZED CIRCULATION OF THE WORLD—JULY

GENERAL CIRCULATION

The general circulation of the earth's atmosphere is due to three main causes:

1. The differential heating of the earth's surface.
2. The deflection produced by the earth's rotation.
3. The effects of friction.

These give rise to a fairly simple scheme of average wind conditions, which may be represented in figures 30-a and 30-b. Variations from this simple scheme are numerous and occur continuously, of course, but the general features which it depicts are fairly constant and will be considered briefly. To simplify the discussion of the general circulation, each of the three causes mentioned above will be considered separately.

Differential Heating—The heat supplied to the earth by the sun varies greatly from place to place. The largest quantity is received at the equator and the smallest at the poles. As a result, air is heated at the equator, rises and flows toward the poles. There it sinks back to the surface and returns toward the equator (figure 31-a). This is the type of circulation that would occur on a stationary sphere.

The conditions on the earth which tend to cause this type of circulation are shown in figure 31-b, which represents a meridional cross-section of the atmosphere from the equator to the north pole. Average conditions are shown for February for the troposphere and a portion of the stratosphere. By applying the *circulation principle* the tendency to establish a closed circulation from the equator to the north pole, with a northward moving current aloft and a southward moving current near the surface is readily seen. It should also be noted from figure 31-b that the temperature distribution within the stratosphere favors a circulation in the opposite direction, with a flow from the equator toward the poles near the tropopause and a return flow at high levels. This purely thermal circulation also gives rise to ascending currents over the warm equatorial region and to descending currents over the poles. Similar conditions exist in the southern hemisphere.

Deviating Forces—A particle of air which is set in motion by the general thermal circulation is acted upon by a number of forces which tend to deviate it from its initial path. Only two of these forces are of sufficient magnitude to be considered here, namely, the Coriolis force and frictional forces. Several other very minor forces act upon air particles. These will be mentioned here for the sake of completeness and because their importance has frequently been overestimated.

First, there is a slight difference in the density of two bodies of air, one traveling in the direction of the earth's rotation, and one traveling in the opposite direction. This effect is due to centrifugal

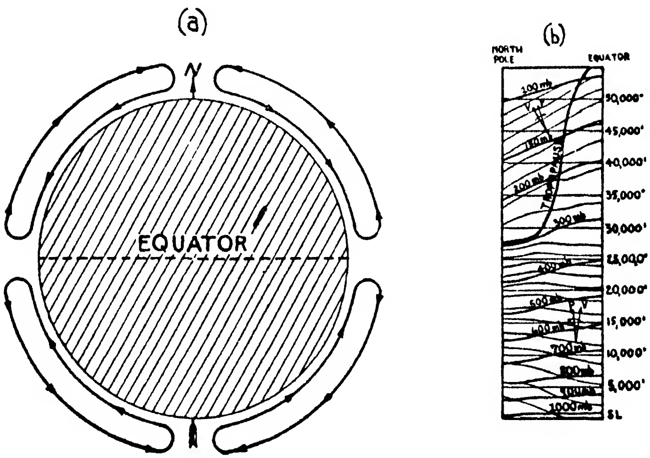

FIGURE 31.—(*a*) HYPOTHETICAL CIRCULATION ON A NON-ROTATING SPHERE. (*b*) SOLENOIDAL DISTRIBUTION ON THE NORTHERN HEMISPHERE DURING FEBRUARY

force, and tends to cause the underrunning of one body of air by the other. Actually this effect is practically negligible.

Then, there is the so-called *conservation of angular momentum* effect which might appear, upon casual consideration, to be rather important. A particle with a certain angular momentum ($mr^2\omega$) tends to retain that momentum as its radius is varied. Thus, if a particle at the equator moves northward along the surface, its distance from the earth's axis will decrease and its radius of curvature will become smaller. In order to conserve the angular momentum, its angular velocity (ω) should therefore *increase*. Theoretically a particle which is at rest at latitude 30° would acquire a velocity of

over 1000 miles per hour if it were forced to latitude 60°. No such velocities, however, are ever encountered on the earth's surface, and it has been shown in various ways that the effect of the conservation of angular momentum is rendered unimportant by frictional forces, and by pressure gradients.

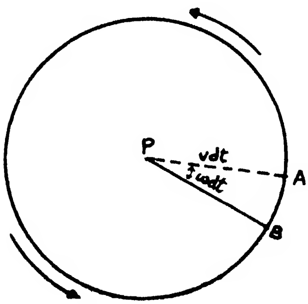

FIG. 32.—ILLUSTRATION OF THE ACTION OF THE CORIOLIS, OR DEVIATING FORCE, DUE TO THE EARTH'S ROTATION

The *Coriolis force,* however, is of considerable importance since it acts upon every particle of air as it moves over the earth's surface. This force, in its general form, is rather difficult to explain without the use of spherical co-ordinates. A special case in which a particle at one of the poles is considered, has been used by Humphreys to demonstrate the underlying principle, however.

Let a particle, P at the north pole move freely in the initial direction PA at the velocity v (figure 32). By the time it has arrived at the point in space representing A, the earth will have revolved a distance BA, and the particle will actually reach B, having followed the path PB. If the earth is revolving about its axis with a velocity, ω, the angle BPA will be ωdt after a time dt, and the distance PA will be vdt. Also, let $AB = ds$. For a smaller angle, the angle may be taken to be equal to its sine, or,

1. $\omega dt = \dfrac{ds}{vdt}$

2. $ds = v\omega(dt)^2$

If the initial velocity is zero,

3. $ds = \frac{1}{2}\, a\, (dt)^2$

Equating 2 and 3.

4. $a = 2v\omega$

Since the force producing an acceleration is,

5. $F = ma$

the force producing the acceleration in this case is, substituting 4. in 5., and calling the Coriolis force, d,

6. $d = 2mv\omega$

The component of ω, the angular velocity of the earth at any latitude ϕ, is $\omega \sin \phi$, therefore, at any altitude,

7. $d = 2mv\omega \sin \phi$

From this equation, which may be shown to be true for *all* cases of a particle moving freely over the earth's surface, it is seen that the accelerating force, or *Coriolis force,* is:

1. Directly proportional to the mass.
2. Directly proportional to the straight line velocity.
3. Directly proportional to the angular velocity of the earth.
4. Directly proportional to the sine of the latitude.
5. Always the same, no matter what the horizontal direction of the particle may be.
6. At right angles to the direction of motion of the particle, and therefore it can only influence its direction, never its velocity.

In the northern hemisphere the Coriolis force causes a particle to be deflected to the right, and in the southern hemisphere, to the left. At the equator, no deflection occurs.

Frictional Forces—The effects of uneven heating, and of deflection due to the earth's rotation, on the general circulation have already been noted. There is still a third important force acting upon the atmosphere which modifies the effects of the other two. This is friction. Because of it, winds on the earth's surface do not attain the values which would be expected from the pressure distribution. Furthermore, the surface winds do not blow parallel to the isobars but diverge from them by a considerable amount. Frictional effects are the cause of both of these phenomena, and the latter one, that causing the airflow to deviate from its normal course, is very important in maintaining the circulation over the earth's surface.

The friction between air currents and the surface of the earth, and also between adjacent portions of the atmosphere gives rise to a turbulent zone ranging from a few feet to several hundred feet in width. Throughout this zone a constant interchange of properties

between the earth's surface and the air above it, or between two adjacent bodies of air is constantly taking place. The properties which are commonly involved in this zone of turbulent interchange are heat, energy, water vapor, momentum and various solid impurities. The degree of turbulence within this boundary zone is measured by the exchange of mass across unit horizontal surface, during unit time. This is called *austausch* from the German word meaning "interchange."

The frictional forces between adjacent bodies of air produce stresses within the boundary zone. These give rise to turbulence and cause an interchange of properties between the two bodies of air. Similarly the friction between the earth's surface and the air above it, causes heating or cooling of the lower portions of the air, and a transfer of moisture and impurities. The degree of turbulence which is produced by frictional forces depends upon both the prevailing temperature gradient and the velocity gradient. It has been found that a certain critical value of the austausch must exist in order to produce definite frictional forces between adjacent air masses, or between the earth's surface and the overlying air. Before this critical value of the austausch is reached, no shearing stresses will be established and no turbulent interchange will occur.

Ekman Spiral—The effects of wind friction on ocean currents have been studied rather extensively by Ekman. His theory has been expanded to airflow over the earth's surface by Hesselberg and Sverdrup. They have calculated the velocity and direction of the wind at various elevations above the surface with a constant pressure gradient. The austausch coefficient varies greatly, from a mean value of about 50 near the surface, to several times this value at an elevation of a few hundred meters. According to Hesselberg and Sverdrug the *gradient wind,* or the wind which a given pressure distribution may be expected to produce, occurs at an elevation of about 600 meters (2000 feet) above the surface. Below and above this level the velocity decreases. The decrease in velocity above 2000 feet may be very slight, since it is frequently overshadowed by the general increase in wind velocity at high levels. The direction of the wind gradually changes between the gradient level and the surface. Near the surface, the shift in direction from the gradient wind may amount to nearly 45°, with the wind blowing across the isobars with that angle.

TABLE 3
WIND VELOCITY AND DIRECTION AT VARIOUS ELEVATIONS

Elevation "z" (meters)	Velocity "C" (meters per second)	Wind Angle "A" (degrees)
5	1.5	46
10	2.9	48
20	5.7	48
50	11.7	59
100	17.9	70
200	21.4	85
400	20.1	91
600	20.0	90
1000	20.0	90

Figure 33 and the accompanying table give the velocity distribution at a time when the pressure gradient is sufficient to maintain a gradient wind velocity of 20 meters per second. *A*, is that angle which the velocity vector at any level, *z* above the surface makes

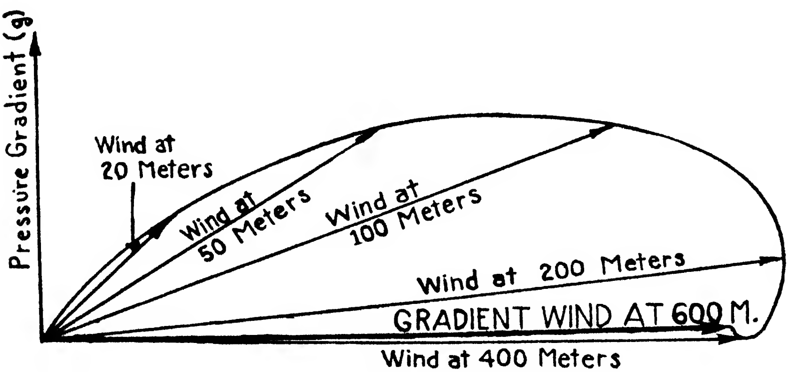

FIGURE 33.—EKMAN WIND SPIRAL FOR GRADIENT WIND OF 20 METERS PER SECOND

with the pressure gradient. It will be noted that the deviation from normal gradient wind flow is the maximum at the surface and that this deviation disappears at the gradient wind level of approximately 600 meters. Ekman demonstrated that if the wind direction and velocity at various levels are represented by vectors

that the ends of the velocity vectors could be joined to form logarithmic spiral. This is the so-called *Ekman spiral.*

The importance of friction is thus considerable, since without it—(1) No circulation could exist, (2) There would be no airflow from high pressure toward low pressure areas, (3) There would be no interchange of properties between the earth's surface and the overlying air or between adjacent bodies of air, (4) The horizontal pressure gradients would reach their maximum at the earth's surface instead of in the free atmosphere.

Gradient Wind—It was shown on page 77 that the Coriolis force, *d*, is equal to,

$$d = 2mv\omega \sin \phi \tag{1}$$

For frictionless horizontal rectilinear accelerated flow, assuming v positive north, and u positive east,

$$\frac{du}{dt} = -2v\omega \sin \phi - \frac{1}{\rho}\frac{\partial p}{\partial x} \tag{2}$$

$$\frac{dv}{dt} = 2u\omega \sin \phi - \frac{1}{\rho}\frac{\partial p}{\partial y} \tag{3}$$

let $2\omega \sin \phi = l$, since it is a constant for any latitude, and also assume that a steady state has been reached so that,

$$\frac{du}{dt} = 0, \text{ and } \frac{dv}{dt} = 0, \quad \text{then} \tag{4}$$

$$lv = -\frac{1}{\rho}\frac{\partial p}{\partial x} \tag{5}$$

$$lu = \frac{1}{\rho}\frac{\partial p}{\partial y} \tag{6}$$

squaring and adding the above equations,

$$l^2(u^2 + v^2) = \left(\frac{1}{\rho}\right)^2\left[\left(\frac{\partial p}{\partial x}\right)^2 + \left(\frac{\partial p}{\partial y}\right)^2\right] \tag{7}$$

taking the square root of both sides,

$$l\sqrt{u^2 + v^2} = \frac{1}{\rho}\sqrt{\left(\frac{\partial p}{\partial x}\right)^2 + \left(\frac{\partial p}{\partial y}\right)^2} \tag{8}$$

now,

$$\sqrt{u^2 + v^2} = c, \tag{9}$$

or the resultant velocity, and

$$\sqrt{\left(\frac{\partial p}{\partial x}\right)^2 + \left(\frac{\partial p}{\partial y}\right)^2} = \frac{dp}{dn}, \text{ or the } \textit{pressure gradient}, \tag{10}$$

thus,

$$lc\rho = \frac{dp}{dn} \quad \text{or,} \tag{11}$$

$$2\omega \sin \phi \, c\rho = \frac{dp}{dn} \tag{12}$$

This states the important fact that for the steady state, *the horizontal pressure gradient is just balanced by the Coriolis force.*

The velocity of the wind for which the deflective force due to the rotation of the earth is just balanced by the horizontal pressure gradient is termed the *gradient wind*. This is illustrated in figure 34 (top) which shows straight line flow for the northern hemisphere. The pressure gradient, g, acts toward the region of low pressure to the left. The Coriolis force, d, acts to the right of the velocity vector, c, which represents the gradient wind. The value of c is obtained from,

$$c = \frac{dp}{dn} \cdot \frac{1}{2\rho\omega \sin \phi} \tag{13}$$

For curvilinear flow, an additional force, the *centrifugal force* must be considered. This acts outward from the center of rotation. It is equal to,

$$f = \frac{v^2}{r} \tag{14}$$

where f is the centrifugal force, v the velocity of rotation, and r the radius of curvature.

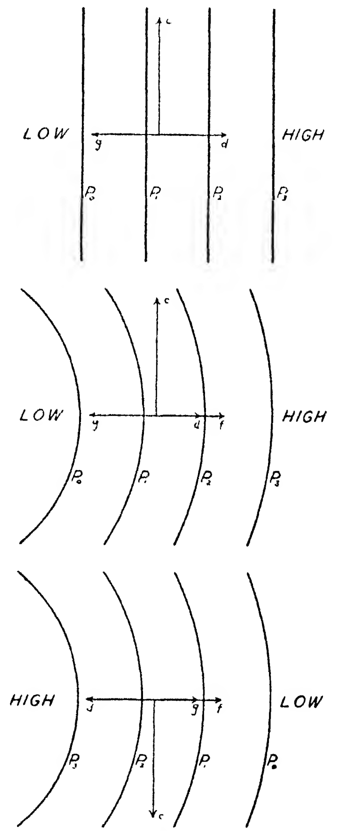

FIGURE 34.—GRADIENT WIND RELATIONSHIPS IN THE NORTHERN HEMISPHERE

Top, straightline flow; center, cyclonic flow; bottom, anticyclonic flow.

In figure 34 (center) the conditions for steady flow for *cyclonic* circulation in the northern hemisphere are shown. It will be noted that the pressure gradient, g, is balanced by the sum of the Coriolis force, d, and the centrifugal force, f.

Conditions for *anticyclonic* circulation in the northern hemisphere are shown in figure 34 (bottom). Here the pressure gradient is balanced by the *difference* of the Coriolis and the centrifugal force.

For *cyclonic* flow the gradient velocity is given by,

$$c = \sqrt{\frac{r}{\rho}\frac{dp}{dn} + (r\omega \sin \phi)^2} - r\omega \sin \phi \tag{15}$$

For *anticyclonic* flow the gradient velocity is given by,

$$c = r\omega \sin \phi - \sqrt{(r\omega \sin \phi)^2 - \frac{r}{\rho}\frac{dp}{dn}} \tag{16}$$

With cyclonic flow it is seen that the two terms under the radical are additive. There is therefore no limit to the velocities that can be developed under these conditions. With anticyclonic flow, however, the two terms under the radical are subtractive, and there is a definite limit to the maximum attainable velocity. Actually the maximum velocity with anticyclonic flow is,

$$c_{\max} = r\omega \sin \phi \tag{17}$$

Table 24 in the Appendix gives values for gradient winds for various pressure gradients at various latitudes for both cyclonic and anticyclonic circulation. This table may be used for the construction of scales to determine the value of the gradient wind from the spacing of the isobars. Such scales are very useful in determining the approximate wind velocity in regions where surface winds are unreliable, as over high mountain areas, or where they are unavailable, as over the ocean. Such determinations of the value of the gradient wind may be used indirectly for estimating the rate of movement of fronts. It has been found from experience that over the open ocean, fronts generally move with a velocity of 70% to 80% that of the gradient velocity and over relatively flat land surfaces, that they move with speeds of 50% to 60% that of the gradient velocity.

CIRCULATION OF THE TROPICS AND SUBTROPICS

Due to differential heating between the equator and the regions just north and south of it, the circulation here is established according to the circulation principle. Air at the surface near the equator

rises and moves north or south, while air near the surface flows toward the equator. The deflective effect of the earth's rotation, or Coriolis force, causes the upper portion of this circulation to be changed to great eastward moving currents as the air moves away from the equatorial zone. This important branch of the general circulation is known as the *antitrade* wind and occurs at elevations of several thousand feet above the subtropics as a strong west wind. As these great currents of air move north and south away from the equator, they are gradually cooled until at approximately 30° latitude they sink to the earth's surface and appear in the high pressure belt which encircles the earth at these latitudes as comparatively feeble westerly surface winds. This region of general high pressure which is characterized by descending westerly air currents, is generally known as the *Horse Latitudes.*

This general westerly current splits in the Horse Latitudes. A portion of it returns southward, while another portion continues northward in the zone of prevailing westerlies. This northward branch will be discussed in the section on the circulation of middle latitudes. The portion which returns southward toward the equator completes the closed link of the general subtropical circulation. This portion of the earth's atmosphere contains what may be called a *direct circulation*, since it is produced by simple thermodynamics, as opposed to the *forced circulation* of the middle latitudes. The southerly moving current of this direct circulation is also affected by the Coriolis force. It is deflected to form, in both hemispheres, the very important *trade wind belt* of strong easterly winds.

Within the general region of the tropics and subtropics there is thus found at the surface, near the equator, a region of comparative calms associated with rising air currents accompanied by a general belt of low pressure. Farther to the north and south this zone of light variable winds gives way to the trade wind belt of strong easterly winds. Still farther from the equator, between latitudes 25°–30°, a general high pressure region associated with descending air currents, and light northeasterly to northwesterly winds is encountered.

In the upper levels, the winds near the equator are also light and variable, becoming strong westerly currents (the antitrades), to the north and south of the equator. Farther to the north and south in the region of the Horse Latitudes they descend to the surface to form the westerly surface winds of this region.

CIRCULATION OF THE HIGH LATITUDES

Circulation in the polar regions is similar to that in the tropics and subtropics in that it is a thermodynamically direct one. Here there is a general cooling and sinking of the air at the poles. This spreads out near the surface toward the equator into a current which is deflected by the Coriolis force into moderate easterly winds. As this current reaches approximately latitude 60° the general equatorward movement disappears and a portion of the current continues southward into the zone of prevailing westerlies, while another portion returns aloft toward the poles to complete the direct thermal circulation of this region.

It is thus found that the polar regions are characterized in general by relatively high pressure together with rather light variable winds which have a marked descending component, and that the subpolar regions between latitudes 60°–80° are occupied by moderate northeasterly and easterly winds. The southern portion of this general subpolar circulation is marked by ascending currents, and a general belt of low pressure in which lie the important, permanent low pressure areas such as the Aleutian and Icelandic LOWS.

CIRCULATION OF THE MIDDLE LATITUDES

The southern portion of the subpolar region at about latitude 60° represents an important discontinuity in both wind direction and temperature. At this point the easterly currents of the subpolar region contrast sharply with the general westerly circulation of the middle latitudes. Also the temperature of the cold air which is moving away from the poles contrasts strongly with the relatively warm air of equatorial origin which has come from the equator, after descending to the surface in the Horse Latitudes.

Since the average air movement within the middle latitudes has a slight average northward component, it has been a puzzling problem for some time as to how a compensating flow can occur to counteract this continued northward movement. It has been pointed out recently that the only way in which this compensating flow can occur is by means of irregular strong outbreaks of the cold polar air. These rapidly pass southward into the zone of prevailing westerlies and effect a return flow to compensate for the average slight northerly component within this region.

These outbreaks are frequently of great depth, involving practically the entire troposphere, and effect the transportation of very great masses of air at the times they occur. Since no southerly currents of comparable magnitude are ever observed, it is believed that this explains satisfactorily the presence in the middle latitudes of an average northward moving flow involving the entire troposphere. The manner in which these great outbreaks of polar air occur is still imperfectly understood, but it is generally believed that the *polar front* or boundary between the subpolar easterlies and the prevailing westerlies, acts as a barrier to the outflow of polar air under normal conditions. Only when a considerable mass of cold air has built up in the subpolar regions, does it attain sufficient force to break down the barrier and to flow into the middle latitudes. These sudden outbreaks usually occur when a wave disturbance is generated along the *polar front*, generally in the region of one of the semi-permanent areas of low pressure, such as the Aleutian LOW.

The question as to what initiates the formation of these waves has not been answered as yet, although Exner's *barrier theory* is apparently the most satisfactory explanation. According to this theory a small tongue of polar air is projected into the zone of prevailing westerlies along the polar front as illustrated in figure 35. This acts as a barrier to the normal flow in the region near the front and appears to be a plausible explanation for the origin of such waves. Certain investigations which have been carried on very recently by B. Holzman in connection with the study of extremely detailed phenomena along fronts seem to confirm Exner's general theory. Holzman has shown that small protuberances along a cold front may rapidly increase in intensity and give rise to wave disturbances, which then increase rapidly in magnitude. It is believed that this process acts on a large scale in the case of wave disturbances along the polar front, and explains the origin of many of the most important extratropical cyclones.

STRATOSPHERE CIRCULATION

It has been mentioned above, in connection with the general circulation, that the movement of air within the stratosphere is the reverse of that in the troposphere. In other words, the heat transport in the stratosphere is from the poles toward the equator since

the lowest temperatures appear above the equator. In the lower portion of the stratosphere, near the tropopause, wind velocities reach their maximum and it might be expected that considerable air might be transported within this region. However, the density of the air is so low in the stratosphere that no appreciable mass transport can occur. In general the circulation of the northern hemisphere

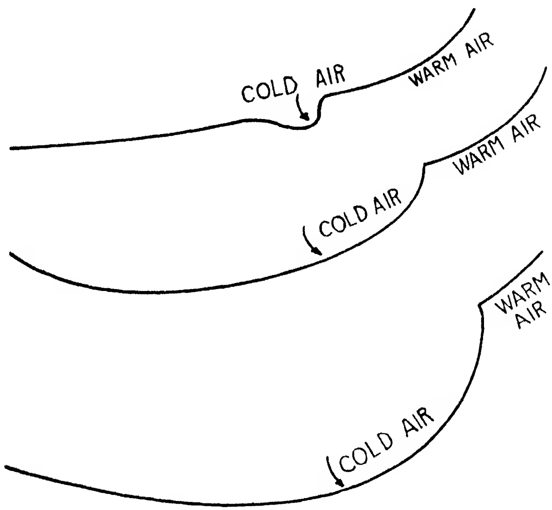

FIGURE 35.—EXNER'S BARRIER THEORY OF CYCLONE FORMATION

The tongue of cold air breaks the smooth frontal surface and initiates the formation of a wave.

is southwesterly in the lower stratosphere and northwesterly in the upper stratosphere, with rising currents over the poles and descending ones over the equator. Comparatively little is known about the details of this circulation due to insufficient data. It is known rather definitely, however, that the high wind velocities frequently ascribed to the stratosphere do not exist, but rather that a gradual decrease in velocity occurs above the tropopause.

CELLULAR CIRCULATION

Bergeron has investigated theoretically the larger aspects of the general circulation and has found that they are best explained by considering that the major circulations within each hemisphere occur in a series of cells. He shows that four high pressure cells are

located in a belt around latitude 30° and four low pressure cells around latitude 60°. This is shown schematically in figure 36 which represents a polar projection of the average field of pressure over the northern hemisphere. Bergeron shows through theoretical reasoning that the point of highest pressure within the individual high pressure cells should be located somewhat to the east of the centers of the cells. This is well borne out by a study of the mean atmospheric pressure of the earth's surface in which the centers of the important high pressure areas of the northern hemisphere, the Azores and the Pacific HIGHS, are located 30° to 40° east of the centers of the cells in which they occur.

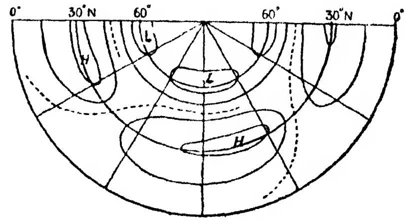

FIGURE 36.—SCHEMATIC PRESSURE FIELD OF PART OF THE NORTHERN HEMISPHERE

This shows the subtropical high-pressure belt broken up into several "cells"; also the subpolar low-pressure belt. The intersections of the cells at the surface are marked by the dashed lines.

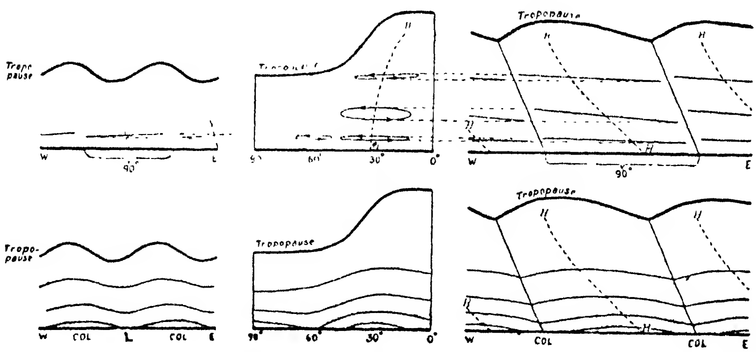

FIGURE 37.—GENERALIZED CIRCULATION OF THE TROPOSPHERE

Upper figures: trajectories of particles; lower figures: generalized isobars. Left: zonal section along latitude 60° north; middle: meridional section; right: zonal section along latitude 30° north. Note the tendency of the axes of HIGHS to slope from east to west aloft; the manner in which the tropopause reflects the surface pressure; the tendency of particles to sink as they proceed toward the east in the subtropical cells.

Bergeron also shows theoretically that the trajectories of particles within these cells are ellipses, which are greatly elongated in a north and south direction as shown in figure 37. Furthermore, as may be noted from the right hand portion of figure 37 the ellipses which represent the trajectories of the air particles are oriented in

such a manner that particles of air to the south of the anticyclones tend to rise as they move toward the west, and to sink as they move to the east. The inclination of the trajectories is probably about one in ten thousand. This inclination would give particles a vertical amplitude of about one kilometer as they pass from one side of a cell to the other along a parallel of latitude. The strongest inclination of the trajectories occurs at an elevation of approximately 5 kilometers above the surface. Above this elevation the inclination again decreases. The upper sides of the cells are apparently limited by the tropopause as shown in figure 37.

The boundaries between the cells are marked by discontinuity surfaces which have a slope of about one to two hundred. The intersections of these cell walls with the surface are marked by regions of somewhat lower pressure which may be called *cols*. To conform with this idea, the tropopause is indicated in figure 37, as being somewhat lower above these cols and somewhat higher over the center of the cells. It will be noted that the lines of highest pressure are not vertical, but rather slope from east to west, so that the crests of the HIGHS along the tropopause are a considerable distance west of their location at the surface.

Bergeron points out that although it is impossible to show directly the general rising tendency of air as it passes from the east to the west along the southern borders of the anticyclonic cells, yet this is shown indirectly by general climatic conditions. It is found that in the western portions of the anticyclonic cells, where ascending currents are the rule, heavy rainfall prevails, while in the eastern portions of the anticyclones, desert conditions are found. Thus the western portion of the Azores HIGH includes the Gulf of Mexico and Caribbean Sea area together with the southeastern coast of the United States. All of this region is characterized by relatively heavy rainfall. On the other hand the eastern portion of the Azores HIGH includes the Sahara Desert and the relatively dry area of southwestern Europe. According to this theory, the heavy rainfall in the western portion of the HIGH is probably due in part to the gradual ascent of the air in the southern portion of the Azores High Pressure cell as it moves westward, with a marked tendency to become convectively unstable.

DETAILS OF THE GENERAL CIRCULATION

V. Bjerknes has developed a scheme of the general circulation of the northern hemisphere which is very useful in determining the position of regions of abnormal and subnormal atmospheric activity. Reference to figure 38 shows a number of features of interest in connection with the various circulations which have been discussed. Of particular interest is the polar front, which is shown both in plan and cross-section. In plan is seen a series of wave disturbances, each of which is an individual cyclone. The largest of these disturbances occupies the entire zone of prevailing westerlies and in fact

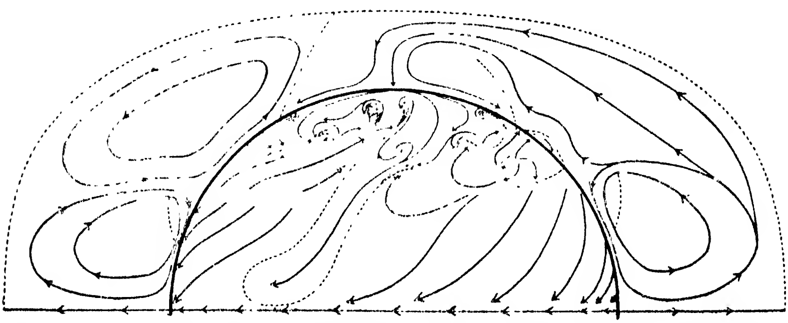

FIGURE 38.—GENERALIZED ATMOSPHERIC CIRCULATION OF THE NORTHERN HEMISPHERE (AFTER V. BJERKNES)

Note the Polar Front with a number of wave cyclones in various stages of development the ascending air over the equator and descending air in the "Horse Latitudes" (30°–40° north); easterly winds in the Trade Wind belt, westerly winds in the middle latitudes, and easterly winds in the high latitudes; the "Intertropical Front" between the Trades and the Antitrades; and the tongue of Polar air extending as a continuous channel well into the tropics.

extends its influence into the subtropics. The smaller ones are confined to the northern portion of the middle latitudes. This transport of large masses of air thus explains very clearly the exchange of air between the subpolar and the temperate zones as brought out on page 85.

In cross-section may be seen the polar front surface occupying a major portion of the troposphere. It lies between the zone of prevailing westerlies to the south and the subpolar belt of easterly winds to the north. The marked convergence in surface airflow is to be noted here.

The contrast between the trade winds and antitrades in the subtropics is also clearly shown in the sectional view of this figure.

Actually the discontinuity between these two currents is sufficient to make it a major atmospheric discontinuity.

Figure 39 represents a polar projection of the northern hemisphere centered on the north pole. The principal regions of high and low pressure are indicated together with the major areas of air convergence. These regions of convergent airflow are important, since they represent the source of a large number of the major storms which occur in this hemisphere. The general high pressure belts of the northern portion of the subtropics with their descending airflow and stagnant winds are shown. The principal centers of action of the middle latitudes are also indicated, including the two great high pressure areas lying over Canada and Siberia with the intervening low pressure regions of Iceland and the Aleutian Islands.

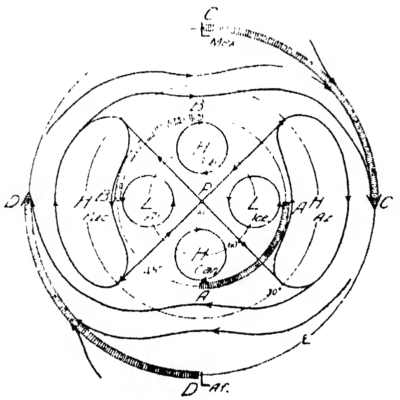

FIGURE 39.—BERGERON'S SCHEME OF THE GENERAL CIRCULATION OF THE NORTHERN HEMISPHERE

Note the semi-permanent Icelandic and Aleutian LOWS; the Canadian and Siberian HIGHS, and the Pacific and Azores subtropical HIGHS. Regions of convergent airflow are shown as heavy shaded lines.

These various high and low pressure regions form a series of closed links within the general circulation. Along their boundaries are located two major regions of convergent airflow. These areas of convergence are the source of most of the important waves which occur along the polar front. The intervening regions of divergence represent areas of decaying storms. Thus, it will be seen that the principal regions of frontal formation are in the China Sea, and along the eastern coast of North America, while regions of frontal disintegration are the west coast of North America and northwestern Europe. This information is useful to the forecaster in determining where the formation of new cyclone families is most likely to occur.

Figures 40 and 41 are examples of detailed streamline maps. They show regions of convergence and divergence, as well as areas with various average velocities. This type of chart is useful in frontogenesis studies.

MINOR CIRCULATIONS

Monsoon Winds—One of the most important of the minor circulations of the atmosphere is that caused by differential heating of adjacent portions of the earth's surface. The *Monsoon* wind of India is the best illustration of this type and has been subjected to much detailed study. Here the temperature contrast between the interior of India and the Indian Ocean is very considerable and gives rise to seasonal winds which greatly influence the climate of the

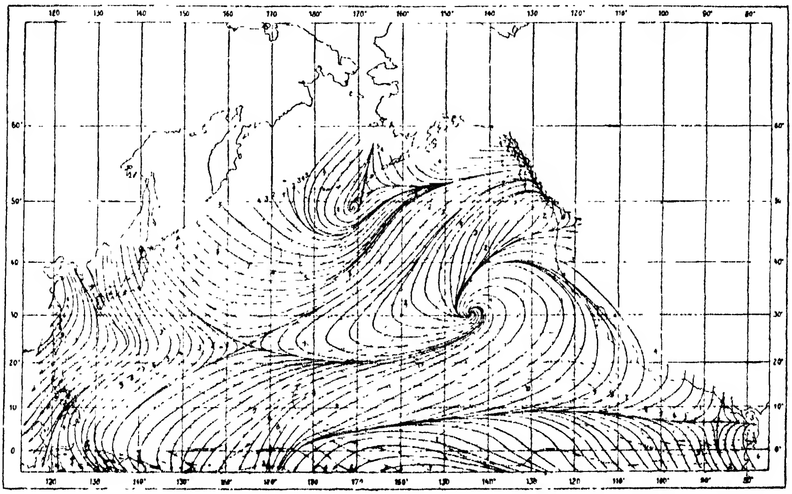

FIGURE 40.—CIRCULATION OVER THE NORTH PACIFIC OCEAN DURING JANUARY

The solid lines are streamlines, and the light dashed lines are curves of equal velocity. The small numerals give the velocity in meters per second. The subtropical HIGH is located near 30° north and 145° west. It is represented by a center of *divergence* of the streamlines. The Aleutian LOW is located near 50° north and 170° west and represents a center of *convergence* of the streamlines. The "equatorial line of convergence," located between the equator and 10° north represents the boundary between the southeast and the northeast trade winds. (After Werenskiold.)

entire Indian peninsula. In the summer, when the continent is highly heated and the Indian Ocean remains relatively cool, a strong breeze is produced, blowing from the Indian Ocean toward the interior of the continent with a return circulation at higher levels. This is a thermodynamically direct circulation of a type which is very common throughout the world. Its counterpart is found on a similar scale along the margins of other continents, and on a smaller scale in the well known land and sea breeze, as well as in other

thermally produced winds. These are all explained by an application of the Circulation Principle as shown in figure 29-c.

Gravity Winds—Many other types of local winds which do not exhibit the closed circulation of the Monsoon are due to ordinary gravitational effects. Examples of these are the *Bora* which flows down the margins of high plateau regions, to adjacent low lying coastal areas. This is generally produced in the winter when an extensive anticyclone occupies the plateau and produces extremely cold air. By virtue of its high density this flows through passes in

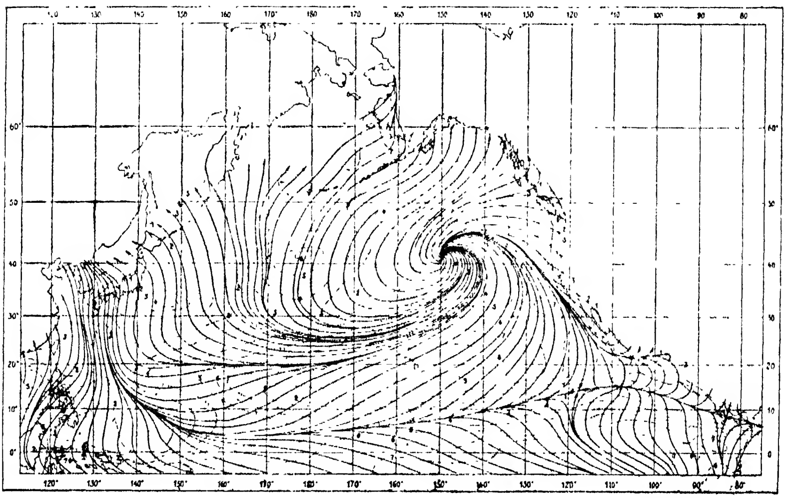

FIGURE 41.—CIRCULATION OVER THE NORTH PACIFIC OCEAN DURING JULY

The subtropical HIGH is seen to be located near 40° north and 150° west, and represents a center of divergence of streamlines in that vicinity. The "equatorial line of convergence" runs in a general east-west direction near latitude 10° north. It represents marked convergence of the streamlines. The east Asiatic summer monsoon is strongly developed at this time of the year as shown by the strong onshore current in this region. The Aleutian LOW has disappeared. (After Werenskiold.)

the margins of the plateau with very high velocities. In some cases the Bora may also exhibit certain characteristics of the Monsoon, although insufficient investigations have been carried on to determine whether the returning branch of the circulation exists aloft.

Regions particularly affected by this type of wind are the coasts of Greenland and of Antarctica. In both of these regions the interior of the continental area consists of a great ice cap lying over a high plateau. Along the borders of the plateau, the cold air which accumulates over the ice cap flows toward the ocean with

very high velocities. Mawson reports that during the time he spent in Antarctica during 1912 and 1913 at latitude 67° S. on the edge of the Antarctic continent, that the average wind velocity of an entire year was over 50 miles per hour. Velocities of over 100 miles per hour were frequently recorded with gusts rising to even higher speeds. These winds ceased entirely at distances between one and two hundred miles from shore and were considerably less severe some distance inland where the slope of the plateau is comparatively gentle. Similar reports have been made concerning this type of wind in Greenland where the velocities, particularly in points where valleys reach the sea, attain very high values.

On a smaller scale, this type of wind occurs almost universally, wherever conditions favor the existence of bodies of cold air which can flow to regions of lower air density. Such winds are found at the mouths of caves, at the bases of glaciers, and along the edges of snow fields.

Foehn Winds—Winds which are produced by differences in atmospheric pressure rather than by ordinary thermodynamical causes, are typified by the *Foehn* wind, or *Chinook,* as it is called in North America. This type of wind is caused by descending air whose circulation is established by the regional pressure distribution. Thus, whenever an anticyclonic area occupies a region which is bordered by an area of relatively low pressure lying at a lower elevation, the forced circulation, due to difference in atmospheric pressure, causes a wind to blow from the higher to the lower elevation. As the air flows downward it is heated adiabatically. Since this heating amounts to approximately 5.5° F., per 1000 feet and since the ordinary lapse rate in the atmosphere is only 3°–4° F., per 1000 feet, it is clear that this type of wind may introduce very marked temperature contrasts when it appears. The temperature contrast is especially striking when the foehn wind also results in the influx of a new air mass which is considerably warmer than that which it displaces. This situation is well exemplified by the *Chinook* wind of the western great plains of North America, during which the most remarkable temperature contrasts often occur. H. N. Johnson reports that "—at Havre, Montana, on January 19th, 1892, the temperature rose 43° in 15 minutes—from severe cold (− 6° F.) to mild, balmy air (37° F.) with the arrival of the Chinook." Although this is an unusual case, it is not at all uncommon for the temperature in this general region to rise from 20° to

30° in less than an hour. Johnson reports two cases in which the temperature at Rapid City, South Dakota, rose 67° in 18 hours, both instances occurring during December, 1933.

The climate of the entire western great plains region of Canada and the United States is greatly affected by these Chinook winds. They usually occur with the eastward advance of Polar Pacific air masses which, under favorable pressure distribution, displace the cold Polar Canadian air lying at the surface. The Chinooks often cause very widespread thaws in this region. Occasionally they may even give rise to dust storm conditions.

Foehn winds are to be found, as might be expected from their mode of occurrence, in all portions of the earth's surface which show topographic contrasts. They are very important in modifying the climate of large areas. Particularly interesting in this regard is the climate of southern California. In the winter time the climate here is greatly modified from its normal maritime nature by the frequent occurrence of foehn winds blowing through mountain passes from the interior. In this region the foehn winds are called locally, "Santa Anas," or "desert winds." This latter name indicates that it is believed generally that the warmth and dryness of these winds is due to their desert origin. However, I. P. Krick has shown that the temperature of these winds over the adjoining desert regions at times when they occur, is generally considerably *lower* than that of the coastal areas which they affect, and that their unusual warmth and low humidity are due entirely to their foehn origin. Many other localities along coastal areas which are bordered by mountain ranges are affected by foehn winds, which frequently modify the climate to nearly as great an extent as in southern California. Foehn winds, particularly when they reach high velocities, may be of some importance to aviation since a high degree of turbulence, with strong updrafts and downdrafts, is frequently found in these winds. Pilots generally recognize this fact and avoid the mouths of mountain passes whenever strong winds are blowing.

Foehn Cyclones—In certain cases of pronounced foehn activity it has been found that cyclones may develop due to the marked temperature contrast which arises between the air in the foehn, and that which it is displacing. Such cyclones are occasionally found along the eastern slope of the Rocky Mountains in southern Canada where the temperature contrasts, as pointed out above, may amount to 40°–50° F. Such cyclones may deepen rapidly due to the in-

tense density contrasts present, and may cause very high winds at the surface. Dust storms which frequently affect the northern Great Plains of the United States are often of this origin. They may be expected at any time in the spring when the Great Plains region is occupied by unusually cold Polar Canadian air which is being displaced by warm, dry, Polar Pacific air descending the east slopes of the Rockies as a foehn wind.

Local Convection Currents—The turbulence in the lower atmosphere which is frequently noted on warm afternoons in the summertime over rough topography is due to convectional activity caused by surface heating. This turbulence is frequently present even over flat country, but in this case the vertical currents are never so pronounced as in mountainous regions. This common phenomenon of strong air turbulence over rough country is due to the unequal heating of various portions of the surface. Thus, the mountain tops become considerably warmer than the surrounding air, so that the air in contact with the surface of the ground tends to rise because of its lower density and develop well defined currents. Pilots flying over mountainous regions recognize this phenomenon and expect strong *updrafts* over mountain peaks. Similarly, over shaded valleys and over water surfaces, where the air is cooler than the average temperature of the surrounding atmosphere, there is a tendency for the production of descending currents, which give rise to the well known *downdrafts* often encountered in flying over such areas.

The vertical velocities produced by such thermal currents may be of very great magnitude. Pilots flying over regions of strong surface heating, as over the summits of high mountains, frequently observe updrafts with velocities over 1000 feet per minute. In the case of thunderstorms, which are essentially due to violent vertical air currents, it has been shown beyond doubt, that vertical velocities of over 2000 feet per minute are common and occasionally much larger velocities must exist to support the hailstones frequently found in connection with such activity.

Northers—During the winter season, the Gulf coast and lower Mississippi Valley, together with northern Mexico, occasionally experience marked cold waves accompanied by strong northerly winds. These affect also the neighboring waters of the Gulf of Mexico, and the Pacific Ocean lying off the Mexican coast. They are locally called *Northers* along the Gulf coast, *Tehuantepecers*

when they occur in the Gulf of Tehuantepec, *Papagayos* when they occur in the Gulf of Papagayo along the northwestern coast of Costa Rico, and other local terms. These storms are due to the invasion of these areas by strong outbreaks of polar air which follow cold fronts. They may, therefore, be forecast if the southward progress of the cold front is noted as it approaches these regions. The intensity of the norther depends upon the contrast in the air temperatures on the two sides of the cold front, and on the wind velocities in the cold air mass. Reliable indications as to the speed of the cold front can be obtained from the pressure tendencies in the vicinity of the front. The wind velocities that may be expected can be estimated from the wind velocities both at the surface and aloft within the cold air mass.

As the cold polar air passes over the warm waters of the Gulf, it acquires a great deal of moisture and heat in its lower levels and becomes convectively unstable. When this air reaches the Mexican coast south of Brownsville, it produces very heavy thunderstorms and intense rainfall. As it flows across the Gulf of Mexico and the central part of Mexico and moves southwestward across the Isthmus of Tehuantepec it acquires further intensity where it is confined behind mountain passes. It finally goes down the western slope of the mountain ranges and strikes the Pacific Ocean as a foehn heated gale of terrific force. The concentration of isobaric surfaces frequently reaches its maximum in this region.

The passage of the cold front off the Pacific coast is generally marked by a short interval of squalliness, following which the sky clears and the wind continues to blow under the influence of the general pressure distribution. The occurrence of this type of storm may often be forecast by the appearance of a heavy haze over the entire region. This is a result of the influx of tropical air brought from the Tropical Pacific source region by the pressure distribution preceding the passage of the cold front.

The same type of weather situation explains the strong northers occasionally experienced in Costa Rico and along the west coast of Central America as far south as the Canal Zone. These are invariably associated with vigorous outbreaks of Polar Canadian air and the development of a strong HIGH over the southern or southeastern portion of the United States, commonly with a central pressure of 30.50 inches or more. The LOW associated with these outbreaks generally moves unusually far to the south, often lying south

of Cape Hatteras and with a central pressure frequently below 29.40 inches. With less well developed LOWS which do not move as far to the south, the area affected by the norther is usually restricted to the southern part of the United States or northern Mexico and the northern portion of the Gulf of Mexico.

The norther is a very important meteorological phenomenon in the region in which it occurs since it is occasionally accompanied by widespread damage due to the low temperatures and the high wind velocities which accompany it. The severe thunderstorms and high winds which occur in connection with northers along the eastern Mexican coast south of Brownsville make them very hazardous to aviation.

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1. Foehn Winds of Southern California: I. P. Krick. Gerlands Beit. z. Geoph., v. 39, 1933.

This paper presents an interesting discussion of dynamically heated winds in the Los Angeles region of California.

2. Mean Monthly Air Transport over the North Pacific Ocean: W. Werenskiold. Geof. Publik. v. 2, n. 9, 1922.

This important paper presents a study of circulation over the north Pacific. The author plots streamlines which he obtains from pilot charts of the U. S. Navy. He then investigates the mathematical significance of these streamlines. He points out lines of convergence between various currents of air. Of particular interest is the line of Equatorial Convergence, which represents the breeding ground of many hurricanes.

For detailed information about any particular region, the pilot charts of the U. S. Navy are to be recommended. These are published monthly for the more important regions of the earth and quarterly for the less important areas. They include a vast amount of information regarding the winds, temperature, and ocean currents.

CHAPTER 5

AIR MASSES—GENERAL

INTRODUCTION

The idea of an *air mass* has helped more than any other single concept of meteorology to explain the origin of storms, how they develop, and how they move. It has emphasized the fact that storms usually consist of several different types of air, each having its peculiar characteristics. Furthermore, it has aided the study of the atmosphere in three dimensions. Most important of all, it has given the forecaster a new and useful means of analyzing the synoptic weather chart.

An air mass may be defined as: *an extensive portion of the atmosphere which has everywhere more or less uniform properties at each level.* This means that ideally an air mass should have the same potential temperature and specific humidity everywhere at each level. Actually this is not strictly true, for variations in these properties occur. They are always gradual, however, and *continuous* from one portion of the air mass to another. This characteristic of all air masses of showing *gradual* changes in significant properties is well shown by many of the large bodies of air which enter the United States from central Canada during the winter. The air on the east side of these air masses advances southward, fresh from its source region, retaining its initial coldness and dryness. The air which returns northward on the west side of the anticyclone, however, is considerably warmer and more moist from its journey over the northern and central United States. All of this air is actually of the same type, Polar Canadian. The properties of the fresh air moving southward and the modified air moving northward may differ considerably, yet the change from one to the other is gradual and continuous.

When this air from Canada pushes southward and encounters air in the northern United States which has recently arrived from the Gulf of Mexico, however, the transition from one to the other is very marked and occurs within a very few miles. Frequently the temperature contrast in such cases may amount to 40° F. within

a few miles, and the specific humidity may rise from less than 1 gram per kilogram in the Polar Canadian air to over 10 grams per kilogram in the air from the Gulf of Mexico. These rapid changes in significant properties are never found within individual air masses. They are always the sign of boundaries between distinct types of air.

FRONTS

When air masses move rapidly the boundaries between them (which may be called *Fronts* for convenience), generally remain sharp, and contrasts as sharp as the one pointed out above may often be found. When they move more slowly, though, they tend to inter-mix along their fronts. Under these conditions the fronts become more and more diffuse, until finally the two air masses which were once separated by a sharp boundary may become so thoroughly mixed that it is not possible to identify either one as a separate entity. This variation in the sharpness of fronts makes it difficult at times to locate them on the weather chart, since they may vary in width from about 3 miles in the case of extremely sharp fronts to about 50 miles in the case of diffuse ones. In all cases, however, the front itself will show a decidedly more rapid change in significant properties than the air masses on either side of it. This fact should be remembered and considered carefully in locating fronts on the weather map. It is of especial importance when locating the cold front at the forward edge of a Polar air mass. In this type of air the properties within the air mass vary much more rapidly than in Tropical types, so that there is often a tendency to place the front well within the Polar mass where low temperatures are found. The front is actually located at the forward edge of the advancing body of air where the temperature may seem rather high (due to surface modifications) but where the *rate of change* of temperature and humidity between the Polar and the Tropical masses is the greatest. Details of the structure and motion of fronts are discussed in chapter 9.

AIR MASS PROPERTIES

Whereas the location of fronts at the surface is often rendered difficult due to mixing of the surface strata, their recognition is gen-

erally fairly easy at elevations of several thousand feet above the surface, where turbulent mixing is less pronounced. Even in the case of diffuse surface fronts, it is frequently possible to find marked differences between adjacent air masses when aerographic soundings are available in the frontal regions. By the use of these soundings it is often possible to identify with certainty air masses which have moved long distances from their sources and whose surface layers have been considerably modified.

In identifying air masses it is important to employ properties which are not altered by the ordinary adiabatic processes as the air is raised or lowered in crossing mountain ranges, in rising along fronts, or sinking along surfaces of subsidence. *Specific Humidity, partial potential temperature of the dry air, and equivalent potential temperature* are especially useful in studying air masses. All of them remain unchanged by adiabatic transformations that do not involve the addition or subtraction of water, and the equivalent potential temperature remains unchanged even during condensation and evaporation. Temperature, relative humidity, and other non-conservative properties are of little value in identifying air masses.

AIR MASS DIAGRAMS

A knowledge of the vertical structure of an air mass is of great importance. It not only serves to identify it, but also indicates its stability conditions and moisture content. Various thermodynamical diagrams have been used to describe air masses, among them the Tephigram, the Thetagram and the Rossby diagram (see pages 58-64). The first of these diagrams employs the non-conservative property of *temperature* for one of its coordinates. It is therefore not well suited for use where this element may be affected as the elevation of a layer changes. The Thetagram is useful in showing the variation of the equivalent potential temperature with elevation. Since the equivalent potential temperature is a conservative element, this diagram is of considerable use in studying day-by-day air mass changes at an individual station. It is not widely used in comparing air masses at different localities, however.

The Rossby diagram is of great utility since it employs as coordinates, elements of the weather which are strictly conservative for adiabatic changes which do not involve condensation. The Clark

modification of the Rossby diagram has the slight advantage of spreading out the region of convectional instability over a larger portion of the diagram than the original Rossby diagram. For that reason, it will be used here. The Thetagram will also be used in some cases where its advantage in comparing air mass properties at the same altitude justifies its use. A knowledge of the characteristic curves of various air masses is indispensable to the practical forecaster, and the reader is urged to familiarize himself thoroughly with the appearance of these curves for the various air masses with which he is likely to deal.

AIR MASS NAMES

For the sake of convenience, air masses of different properties which originate in different regions are given distinctive names. These indicate that the air which is being considered has certain very definite properties. They are thus useful to the forecaster since they provide a "shorthand" method of describing air mass properties. Instead of saying that a "cold, dry, stable body of air is present" the meteorologist can say a "Polar Canadian air mass is present," and convey the same thought more concisely.

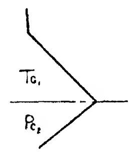


FIGURE 42.—AIR MASS SYMBOL TO SHOW STRUCTURE OF AIR ABOVE A STATION

This is a miniature representation of the equivalent potential temperature diagram.

Actually the most important information about an air mass, from the forecaster's viewpoint, is the temperature and moisture distribution within it. The name itself is useful only insofar as it indicates these properties. Thus, it is more important to the forecaster to know the vertical structure of an air mass than to know its origin and previous history. The latter factors are of importance only as they indicate the vertical structure. Thus, if the forecaster knows what the vertical structure of an air mass is, through aerographic soundings, it is of no importance to know the actual name or past history of the air mass type. Long discussions about the origin, history, and trajectory of an air mass are thus futile if the actual vertical structure is known from aerographic studies.

For this reason it is desirable whenever possible to indicate the vertical structure of the air over a certain region by symbols

to supplement the letters and numerals showing its origin and past history. This may be done by the use of a rough curve which represents in miniature, the equivalent potential temperature diagram, for instance. An example of this method of showing the vertical structure is shown in figure 42. Such a small symbol may also show the presence of frontal surfaces at upper levels, as well as the lifts required to saturate various levels, if this is desired. If such symbols are constructed at each station for which aerographic material is available, a very complete picture of the structure of the upper atmosphere can be presented on the surface weather chart. This method of showing air mass structure is especially desirable in the summer, when air mass differences are less pronounced than in the winter.

SOURCE REGIONS

Large areas of the earth's surface which are relatively uniform with regard to: (1) topography, (2) type of surface, and (3) amount of solar radiation received, are called *source regions.* These regions are ordinarily geographic units containing from one to several million square miles and are thus subcontinental or suboceanic in size. They are usually areas of relatively stagnant air flow.

Three general types of source region surfaces are recognized: land, water, and snow or ice. Each of these has very distinctive effects on the air overlying it, as is brought out in the discussion of various types of air masses. The season of the year and the latitude affect both the surface temperature and the thermal structure of the overlying air. The topography is of rather minor importance in the source region, affecting the structure of only the lowermost layers. It is of great importance, however, in modifying air masses after they leave their source regions.

Two types of source regions may be distinguished, depending on whether or not they are commonly occupied by areas of high atmospheric pressure. Regions of average high pressure are termed *primary source regions.* In these areas the atmosphere remains at rest for a long time, and much of it below the tropopause arrives at equilibrium with the earth's surface. Central Canada, the interior of Siberia, the north Atlantic Ocean from Bermuda to the Azores, are examples of such *primary source regions.*

On the other hand, certain localities which are characterized by

active air flow, modify greatly the lower portions of air masses passing over them. The resulting air may have very different properties from that which it attained over the primary source region. These areas of rapid modification are termed *secondary source regions* and are of almost as much practical importance as the primary source regions. The north Atlantic Ocean east of Newfoundland and the Maritime Provinces, where the lower portions of outbreaks of air from central Canada are modified; the north Pacific Ocean between Siberia and the west coast of North America, where outbreaks of air from Siberia and Alaska are modified; the arid regions of the southwestern United States where several different types may be modified, are all examples of *secondary source regions.* In these areas of moderate circulation, the air does not remain at rest long enough to reach equilibrium with the surface, and as a result only the lower 2000–6000 feet acquires its characteristics. The upper portion of the atmosphere remains essentially unchanged during passage over the secondary source regions.

CLASSIFICATION OF AIR MASSES

In any discussion of air mass types it is necessary to classify them according to some logical scheme based on their physical properties. Bergeron differentiates four zonal types: Arctic (A), Subarctic or Polar (P), Subequatorial or Tropical (T), and Equatorial (E). Each of these may be either Maritime (M) or Continental (C). This is worldwide classification and is mainly used in considering air masses from a broad viewpoint. It is very useful in discussing air masses from widely separated portions of the earth. An air mass from central Canada may thus be compared directly with one from central Siberia, and both would be termed *Polar Continental* (cP) according to Bergeron. Air masses from the region off Cape Horn would be comparable to those originating in the Bering Sea. Both would be termed *Polar Maritime* (mP).

Bergeron indicates air mass modifications as they move away from their source region by suffixing a letter to the air mass symbol. He adds a "W" when the air is warmer than the surface over which it passes, and a "K" when it is colder. Thus, cPK stands for Polar Continental air that has moved away from its source region and is colder than the source over which it is passing. mTW stands for Tropical Maritime air that is warmer than the underlying surface.

Although Bergeron's classification is useful in studying air masses from widely separated regions, it is not as clearly descriptive for actual forecasting use as a local classification which specifies the actual geographic location of the source region. With a local classification, the zonal portion of Bergeron's classification is retained, and a term descriptive of the geography of the actual source region is added. Thus: Polar Canadian, Tropical Atlantic, Tropical Saharan, Polar Siberian, Equatorial Pacific, etc.

Air Mass Modification—Modification in air masses which are described by the local classification may be shown in two ways. Willett attaches the prefix "N" to air masses whose properties are transitional from the original air mass. Thus, NPp indicates transitional Polar Pacific air. This is a satisfactory designation for general forecasting purposes.

A more precise means of indicating modification is that employed by Krick. He places subscript numerals before and after the air mass symbol. These indicate the number of days the air mass has been over water or land since it left its source region. Polar Pacific in its source region would thus be indicated, Pp. After it has left its source and remained over the ocean for 1 day it would be designated, ${}_{1}\mathrm{Pp}$; after 5 days ${}_{5}\mathrm{Pp}$. After the air reaches the land and passes inland, the modification would be shown by a figure *after* the symbol, as, ${}_{5}\mathrm{Pp}_{2}$. To one acquainted with the details of air mass modifications, this symbolization gives much valuable information, since it indicates not only that the air mass is transitional in Willett's sense, but also shows qualitatively the amount and kind of modification that has occurred. Krick also uses the letter "R" to indicate an air mass which is "returning" toward the latitude of its source. Thus, RPp indicates Polar Pacific air which has moved to the south and is again returning northward.

An important fault with Bergeron's method of showing modification, by suffixing "K" (cold) or "W" (warm) to the air mass designator, is that this considers only the portion of the air mass in contact with the ground. Actually, the entire vertical structure of the air mass should be considered in a study of its modification. The character of the air mass may change from "K" to "W" at a given locality as day changes to night in Bergeron's classification, yet obviously this change does not represent a significant change in the structure of the air mass as a whole.

REFERENCES

1. Air Mass Symbols for use on Synoptic Charts: I. P. Krick. Jour. Aero. Sci., v. 2, n. 2, Feb. 1935.

This short paper explains the use of symbols for designating air masses. These symbols include information on the number of days that an air mass has been over land and water surfaces since leaving its source region.

2. Rechtlinien einer dynamichen Klimatologie: T. Bergeron. Meteor. Zeit., Bd. 47, Hf. 7, 1930.

This paper, all in German, was one of the first presentations of a logical air mass classification. It is chiefly interesting from this point of view, since many of the ideas advanced have been modified by later workers.

For other detailed papers on the air masses of various regions see the lists of references of the other chapters on air masses. Many of these individual papers bring out important general points that are applicable to a study of this chapter.

CHAPTER 6

AIR MASSES—NORTH AMERICA

INTRODUCTION

The only study of North American air masses that is at all complete is that of Willett, who has employed the local classification with marked success in this region. Willett points out that the pecularities of the geography of North America cause important differences between the air masses found there and in Europe. As a result, close comparisons between the two continents are not possible. First, the shape of the North American continent precludes the possibility of any important source region for Tropical Continental air mass types. In Europe the extensive regions of northern Africa provide a vast reservoir for this air mass. Also, the distinction between "Tropical" and "Equatorial" air masses is very nearly meaningless in North America. All gradations may be found between Subtropical air (as found, for instance, in returning Polar air which has stagnated over the Bermuda region for several days and which is essentially Tropical in character in its lower portions), and true Equatorial air which reaches the Gulf coast directly from the Caribbean Sea region. The same situation holds true for the distinction between "Polar" and "Arctic" air masses, which may be distinguished in the North American region only with difficulty. As a matter of fact, the source region of all far northern continental air masses is the central portion of Canada, and the southern portion of the Arctic Ocean. In the summer this acts as a "Polar" environment, and in the winter as an "Arctic," so that distinctions between these two types have little meaning here.

Willett distinguishes three Polar air mass types, *Polar Pacific*, *Polar Canadian*, and *Polar Atlantic*, five Tropical types, *Tropical Pacific*, *Tropical Continental*, *Tropical Gulf*, *Tropical Atlantic* and *Superior*. One additional type of Polar air mass has been described by Krick and given the name *Polar Basin*. This will be included in the present classification since it seems to be acceptable as a distinctive air mass type.

Classification of North American Air Masses

Latitude	Surface	Local Source Region	Name of Air Mass	Designator	Time of of Frequent Occurrence	General Classification (Bergeron)
POLAR	CONTINENTAL	Alaska, Canada, Arctic Ocean	Polar Canadian	Pc	Entire Year	cP or cA
		Modified over Southern Canada, U. S., North Atlantic	Modified or Transitional Polar Canadian	wPcl (NPc)	Entire Year	cPK, occasionally cPW over No. Atlantic in Summer
		Great Basin of United States	Polar Basin	PB	October to April	cP
		Modified over Central and Eastern United States	Modified or Transitional Polar Basin	PBl (NPB)	October to April	cPW, cPK (Depending on season)
	MARITIME	North Pacific Ocean	Polar Pacific	PP	Entire Year	mP
		Modified over Western Pacific and North American Continent	Modified or Transitional Polar Pacific	wPPl (NPP)	Entire Year	mPK or mPW (Depending on season)
		Colder Portions of North Atlantic Ocean	Polar Atlantic	PA	Entire Year	mP
		Modified over warm portions of N. Atlantic and Eastern U. S.	Modified or Transitional Polar Atlantic	wPAl (NPA)	Entire Year	mPK
TROPICAL	Continental	Southwestern U. S. and Southern Mexico	Tropical Continental	Tc	April to October	cT
	MARITIME	Pacific Ocean between California and Hawaii between latitudes 20°–35° N.	Tropical Pacific	TP	October to March	mT
		Modified over colder regions of North Pacific	Modified or Transitional Polar Pacific	wTP NTP	October to March	mTW
		Gulf of Mexico and Caribbean Sea	Tropical Gulf	TG	Entire Year	mT or mE
		Modified over U. S. or North Atlantic	Modified or Transitional Tropical Gulf	wTGl (NTG)	Entire Year	mTW or mEW in Winter mTK or mEK over land in Summer mTW or mEW over water in Summer
		Sargasso Sea	Tropical Atlantic	TA	Entire Year	mT or mE
		Modified over U. S or North Atlantic	Modified or Transitional Tropical Atlantic	wTAl (NTA)	Entire Year	mTW or mEW in Winter mTK or mEK over land in Summer mTW or mEK over water in Summer
TROPICAL	Upper Atmosphere	Northeastern Limb of Pacific Anticyclone at elevation above 5000 feet. Subsidence effects pronounced	Superior	S	Entire Year	cT

The modifications of source region properties are indicated here according to Krick's method, using subscripts before and after the air mass designator to show the number of days the air mass has spent over water and land since leaving its source region. Willett's "transitional" designator, "N," may be substituted for these subscripts if desired.

Bergeron's general classification is also given, so that North American air masses may be compared with those in other parts of the world. It will be noted that either cP or cA are indicated for Polar Canadian air masses. The choice of designator here depends on the properties of the individual air masses, as it does in the case of Tropical Gulf and Tropical Atlantic air, where either mT or mE may be used, depending on the properties of the upper portions of the air masses.

POLAR CANADIAN AIR MASSES

The interior of central Canada, from the continental divide on the west to the Hudson Bay on the east, and from the Arctic Ocean on the north to latitude 50°–55° on the south, is the source region for one of the most important air masses in North America, *Polar Canadian.* This is a typical "primary" source region with uniformity of surface and stagnant air conditions much of the time. It is bounded on the west by a great natural barrier against incursions of air from the Pacific—the Rocky Mountain Cordillera. The surfaces of Alaska and the Arctic Ocean exhibit properties similar to those of central Canada during the winter. At this time the entire region from the Bering Sea to Labrador may be considered as the Polar Canadian source region.

Winter Properties of Polar Canadian Air Masses—One of the most important influences in determining the character of Polar Canadian air in the winter is the snow and ice which cover the entire source region, including the Arctic Ocean and the Hudson Bay. This uniformity of surface properties is especially well suited to the production of very cold and stable air. Strong radiational cooling, which occurs near the ground, gives rise to intense temperature inversions. The moisture content is very low, usually less than 1 gram per kilogram, due to the extreme coldness. As a result of the low humidity and extreme stability, Polar Canadian air masses generally exhibit no clouds in their source region. Ice crystal fogs are rather

common at very low temperatures (below −30° F.), but otherwise condensation forms are almost wholly lacking, except when other air masses invade the Polar Canadian source region.

H. Wexler has discussed in detail the formation of Pc air from Polar Maritime air. This interesting process is illustrated in figure 43. The temperature-altitude curve of a mass of maritime air is shown by the solid line. This air has a steep lapse rate, approaching the dry adiabatic, and is well mixed throughout. As it moves over a very cold snow surface, the lower levels cool very rapidly. First, only the lowermost few meters in direct contact with the surface

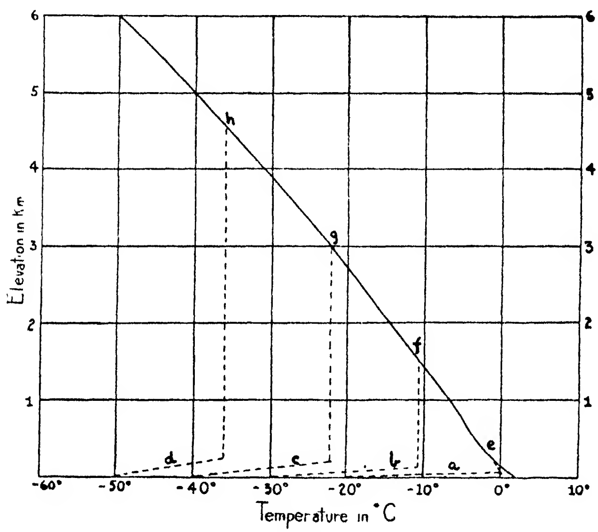


FIGURE 43.—FORMATION OF POLAR CONTINENTAL AIR BY STRONG SURFACE COOLING

cool, as shown by the dashed curve, (a). The higher portions rapidly lose heat to space by radiation, and the surface continues to cool, so that the temperature-altitude curve takes on successively the positions of (b), (c) and (d). The inversion becomes less and less steep, but the thickness of the isothermal layer rapidly increases.

Since this isothermal layer represents the limit of surface cooling, it may be taken to represent the true Polar Continental air. Thus, the top of the inversion, (f), (g) and (h), represents the top of the true Pc air mass. The air above it is still an unmodified Polar Maritime type. The height of the modification caused by either a snow surface (producing Polar Continental air), or a water

surface (producing Polar Maritime air), is rarely over 10,000–15,000 feet. Therefore, the only essential difference in polar air masses is found in the lower levels. Above about 15,000 feet, they are all very similar.

Aerographic soundings from Fairbanks, Alaska, indicate clearly the structure of Pc masses in their source region. These show the intense cold, the dryness, the strong surface inversions, and the isothermal layer. Soundings which have been made somewhat to the south of the actual source region of Polar Canadian air show the same general features. Here (at Ellendale and Fargo, N. D.), surface turbulence mixes the lower thousand feet or so creating a fairly steep lapse rate. Otherwise, there is little difference from the source region properties. The occurrence of a rather steep lapse rate above the characteristic Pc isothermal layer is found here as in the source region. Whether this upper layer is called "Polar Maritime" air, or simply unmodified "Polar" air is unimportant. It is neither true Pc or PM air, but rather an upper level Polar air, perhaps corresponding to the Superior air of lower latitudes.

Polar Canadian air in the source region is occasionally displaced to some extent by warm maritime air from the Pacific. This reaches the surface of central Canada by means of Mackenzie River Valley, and to a much smaller extent through low passes in the Rocky Mountains. The effects of these incursions of Polar Pacific air are often very striking, frequently causing temperature increases in the Mackenzie District of 20°–40° F. These outbreaks of Polar Pacific air are generally caused by the formation of an active wave along the Arctic Front in the general vicinity of Aklavik, Mackenzie, causing a flow of maritime air from the Bering Sea across northern Alaska and the Arctic Ocean. By the time this maritime air reaches central Canada, it has been cooled so greatly (see figure 43) that it is scarcely distinguishable from true Polar Canadian air, and generally joins the circulation of the Canadian HIGH as Pc air.

Occurrence of Pc Air in Winter—Outbreaks of Polar Canadian air are of so many different types that it is impossible to give a synoptic situation that could be called typical of them. In the fall and early winter, when the United States is frequently covered by relatively warm air of Tropical origin, or a greatly modified Polar type, outbreaks of Polar Canadian air generally follow closely on the heels of an old occlusion which has brought Polar Pacific air across the Rocky Mountains into the Mississippi Valley. The cir-

culation established by the old occlusion causes a southerly movement of Pc air from central Canada behind a cold front. As this front moves southward it is usually accompanied by considerable weather activity because of the relatively high humidity and temperature of the air that it is displacing (see figure 84).

When these Pc fronts reach the southern part of the United States they often become practically stationary. At such times they are particularly subject to wave formation. These waves may form singly or in groups of two or three, separated by 500–1000 miles. They frequently cause widespread over-running of the frontal system by warm moist air from the south, with the production of an extensive zone of precipitation and low clouds. At the same time, the low pressure centers which develop at the crests of the waves bring about the incursion of fresh Pc air. This situation is generally terminated by the southward movement of an intense Pc cold front which sweeps everything before it.

In the winter months, outbreaks of intensely cold Pc air which frequently sweep over southern Canada and all of the United States east of the Rocky Mountains apparently do not require the presence of an old occlusion to start their southern movement. Such outbreaks may be expected when (1), the Polar Front moves northward well into Canada and becomes nearly stationary, (2) when relatively warm temperatures occur over most of the Mississippi Valley and eastern coastal area, and (3) when the Canadian HIGH begins to increase in intensity. The first warning of an outbreak of fresh Pc air under these conditions is usually the development of a very slight wave along the Polar Front somewhere between central Alberta and the Hudson Bay. Apparently this slight break in the smooth contour of the front is all that is necessary to initiate the southward movement of a great mass of Pc air, which frequently may cover the entire United States and much of northern Mexico.

At this time of year, waves often form along Pc fronts which become quasi-stationary far to the south, over central Mexico and the Gulf of Mexico. Many of the severe winter storms in Mexico and northern Central America are caused by these waves on Pc cold fronts.

As spring approaches, the Pc outbreaks again resemble those of fall, becoming weaker and weaker as the season advances, with more and more tendency to wave formation over the central part of the United States.

In the winter, outbreaks of Pc air over southern Canada and the United States are generally associated with practically unoccluded cyclones. In the fall and spring, however, the disturbances accompanying these outbreaks are frequently partially, or well occluded. It is important at such times, in making the synoptic analyses, to determine whether the occlusion process is of the warm front type or the cold front type. (See pages 199-208.) While most occlusions are of the latter type, since the Pc air is generally relatively dense, the warm front type of occlusion is more common than ordinarily believed. This is particularly true in the eastern part of the United States, where the general circulation often causes the Pc air to return northward as a relatively warm air mass.

Winter Modifications of Polar Canadian Air Masses—Willett points out that the three most important modifying elements for these air masses after they leave their source region are:

1. The supply of *heat* by contact with, and radiation from, the surface beneath; and the supply of *moisture* by evaporation from this surface.
2. The effect of *subsidence*, which tends to intensify the low level temperature inversions, and to effect a general warming of all the upper strata.
3. The effect of *turbulence*, occasioned by a rough undersurface. This tends to develop a thoroughly mixed lower stratum of air.

The effect of the first of the above influences depends greatly on the type of surface over which the air passes. With snow covered surfaces, modification proceeds much more slowly than with a surface which is free from snow and ice. Thus, at times when the Mississippi Valley region is covered with snow, outbreaks of Pc air reach the southern part of the United States, comparatively little changed. When the ground surface is free from snow, however, the changes in temperature and moisture in the lower strata are often very rapid. This effect becomes more and more pronounced the farther south the air mass travels, since the importance of the sun in warming the surface and inducing evaporation increases rapidly in lower latitudes.

The effect of the type of surface is very striking indeed when the air passes over open water surfaces. Even the rather small

width of the Great Lakes is thus sufficient to raise the temperature of Pc masses passing over them by 15°–20° F. Fresh Pc air passing out over the Gulf of Mexico during the winter, when its trajectory from Canada has been over snow covered terrain, frequently has its surface temperature raised by 40°–50° F. within 36 to 48 hours, and the specific humidity of its surface layers increased from 1–2 gram to 10–15 grams per kilogram. These great changes in the surface properties cause profound modifications in the vertical structure of the air and often give rise to strong convective instability. Northeast winds over Mexico and Central America during the winter thus frequently bring heavy instability showers in Polar Canadian air, which has been highly modified in passing over the Gulf of Mexico.

The effects of large scale subsidence are also very important in modifying the structure of Pc air masses after they leave their source region. Even before they leave central Canada, subsidence probably plays some part in shaping their structure. Willett and Wexler have pointed out, however, that radiational cooling of the surface strata during the intensely cold Arctic nights must account for most of the strong surface inversions which characterize these air masses at their source. As the air moves away from its source region, however, subsidence immediately begins to show its effects. Aerographic ascents thus show conclusively that air from the 4 km level over North Dakota is brought down to about the 2 km level by the time it reaches Oklahoma. This results in warmer and warmer temperatures at lower levels toward the south and a gradual lowering of the elevation of the main inversion. Relative humidities, except at the surface are also lower, level for level, the farther southward the air travels, showing further the effect of subsidence.

The effect of turbulence on Pc masses is of little importance in the Mississippi Valley, since here the land surface is comparatively smooth. It shows its effect markedly however, in those air masses which pass eastward over the Appalachian Mountains. In these cases, the lower few thousand feet become well mixed by turbulence. As a result the surface layers are warmed due to heat which is transported downward from the upper warm portions of the air. Condensation of water in the lower portions of the air as it passes over the mountains also contributes heat to the general warming process. Foehn effects, as the air descends the eastern slope of the mountains, still further tend to warm the lower layers.

Aerographic soundings in the New England region show that the upper levels of Pc air masses which reach that region directly from the northwest, are unusually cold. This is probably due to the fact that these outbreaks are usually associated with deep, low pressure areas over the Maritime Provinces which cause extremely rapid transport of air directly from the center of the source region. The lower portions of these outbreaks, on the other hand, are considerably warmer than similar ones in the north central United States, because they are warmed by the various effects mentioned in the preceding paragraph.

Polar Canadian air only rarely passes westward over the Rocky Mountains into the southwestern portion of the United States, since the stability of the air is too great to permit its ready passage across this barrier. When it does arrive in the Great Basin region, as occasionally occurs in the case of strong outbreaks, however, its characteristics are interesting, in demonstrating the effects of topography. In such cases the air is considerably warmed as a result of turbulent mixing in the lower layers. It is also usually greatly affected by subsidence. On rare occasions, Pc air moves westward through low passes across the Rocky Mountains of southern British Columbia and Washington, and reaches the Pacific coast as a cold wave. The synoptic situation for such outbreaks, is the presence of a low pressure area just off the Pacific coast of Washington or Oregon, together with an intense HIGH over central Canada. The combination of these circulations causes rapid flow of fresh Pc air to the coastal region. Here it produces unusual weather phenomena, such as sub-freezing temperatures with snow flurries, along the Washington and Oregon coasts. In exceptional cases this air may be transported as far south as southern California where it may even cause snow flurries on the coastal plain. The remarkable snowfall in Los Angeles on January 15th, 1932, was of this type.

Occasionally, the average centers of action along the Polar Front may shift 800–1000 miles west of their normal position. The boundary between Polar Canadian and Polar Pacific air then moves into the western Great Basin, instead of lying along the eastern slope of the Rocky Mountains, as it does normally. The entire Pacific coast region may experience unusually severe winters under this situation due to frequent outbreaks of Pc air. This westward shift of the PP-Pc boundary caused the cold winter of 1936–1937 in the Pacific region.

WINTER FLYING WEATHER IN POLAR CANADIAN AIR MASSES

In fresh Pc air, that has had no maritime trajectory, weather conditions are excellent, with clear skies and unlimited visibilities. The air is smooth due to the great stability and lack of surface heating. As the air becomes modified, however, its properties undergo a great change. The very stability which causes its smoothness also allows pollution of the lower levels by smoke and dust. This results in poor visibilities within a day or so after the initial high velocities which accompany a fresh outbreak, have subsided. Above the surface levels, visibility is excellent.

After passing over a water surface, such as the Great Lakes, Pc air is characterized by stratocumulus clouds and instability flurries that hamper flying considerably. These flurries may persist for several days after the incursion of a fresh mass of Pc air. They often extend from the borders of the Great Lakes southward for several hundred miles, and eastward across the Appalachian Mountains. Ceilings and visibilities frequently become very low in these squalls, especially over the rolling country of western Pennsylvania and New York.

The tops of cloud decks under these conditions commonly extend to only 5000–8000 feet over the flat country of Ohio, Indiana and southern Ontario, but they frequently build up to 10,000–15,000 feet over the mountains of New England. This tendency of squalliness to increase over high country is well displayed in the mountains near Albany, New York. It is almost invariably found here that the most unfavorable flying weather in connection with Pc outbreaks occurs directly east of the Mohawk Valley. Air which has passed across Lake Ontario flows along this pathway with practically no resistance until it strikes the west slope of the mountains east of Albany. Here it is forced upward with the production of severe squalliness. North and south of this region, along the east side of the Hudson River, air arriving from the west is first forced to ascend the Adirondack or Catskill Mountains, and thus loses much of its heaviest squalliness. Pilots recognize this fact, and many times find flyable weather on the Albany-Boston route several miles south of the direct course.

Icing occurs in Pc air only after it has had a maritime trajectory. It may then be locally severe. This is particularly true in Pc

Typical Pc Air in Winter
Fairbanks, Alaska

Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A	W g/kg.	Lift Meters
Surface	−41	−42	−55	44	233	0.1	
1000	−16	+ 3	....	40	270	0.5	1300
2000	−16	+ 3	....	30	277	0.5	1400
3000	−21	− 6	....	27	280	0.3	1500
4000	−28	−18	....	24	284	0.1	2200
5000	−35	−31	....	27	288	0.1	1500

Average Pc_{0-1} Air in Winter
Ellendale, North Dakota

Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (444)	−26	−15	−19	82	250	0.3	500
1000	−25	−13	....	82	256	0.4	500
2000	−20	− 4	....	75	272	0.6	600
3000	−22	− 7	....	63	280	0.5	800
4000	−25	−13	....	71	288	0.5	800

Average Pc_{1-3} Air in Winter
Royal Center, Indiana

Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (225)	−23	− 9	−11	91	252	0.5	300
1000	−20	− 4	....	63	262	0.5	700
2000	−16	+ 3	....	40	277	0.5	1400
3000	−18	0	....	77	286	0.9	600

Average Pc_{1-3} Air in Winter
Boston, Massachusetts

Elev. Meters	Temp. ° C.	Temp. ° F	D. P ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (4)	− 6	21	3	43	267	0.9	1400
1000	−14	6	....	50	268	0.6	1200
2000	−18	0	....	50	274	0.5	1200
3000	−23	− 9	....	44	279	0.3	1300
4000	−29	−20	....	48	283	0.2	1200

Average Pc_{2-4} Air in Winter
Broken Arrow, Oklahoma

Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (233)	−15	5	2	86	261	1.0	300
1000	− 9	16	....	60	275	1.2	900
2000	− 8	17	....	50	286	1.2	1200
3000	−10	14	....	45	294	1.0	1300

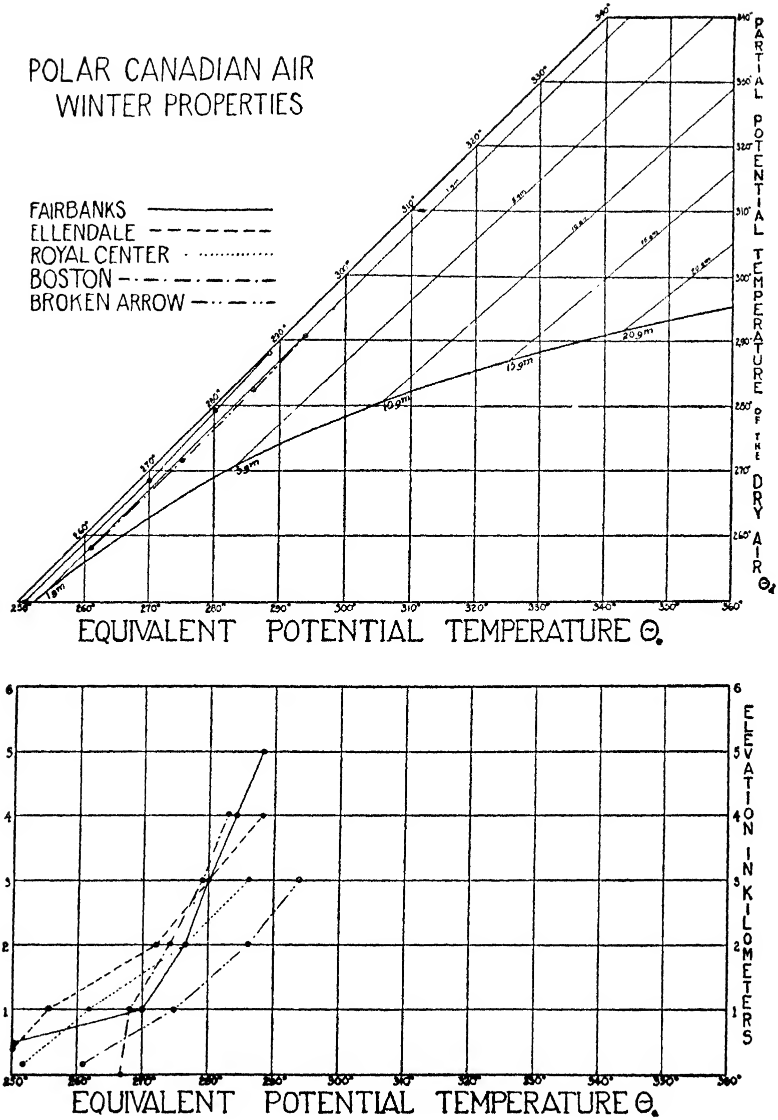

POLAR CANADIAN AIR
WINTER PROPERTIES
FAIRBANKS
ELLENDALE
ROYAL CENTER
BOSTON
BROKEN ARROW
PARTIAL POTENTIAL TEMPERATURE OF THE DRY AIR Θ_d
EQUIVALENT POTENTIAL TEMPERATURE Θ_e
ELEVATION IN KILOMETERS
EQUIVALENT POTENTIAL TEMPERATURE Θ_e

FIGURE 44

air which has passed southward over the Great Lakes. The cold, dry air acquires a considerable amount of moisture as it crosses these water bodies, and surface turbulence, due to high winds, quickly carries this aloft. Here it gives rise to widespread, persistent, stratocumulus cloud decks. These are generally only 1000–3000 feet thick, but they frequently exhibit rather severe icing conditions. This is a result of the very low temperatures, often below 0° F., in connection with the presence of ample water in the liquid state (see chapter 16 for a more detailed discussion of the causes of icing conditions). Pilots often find that they pick up ice very rapidly when passing through these stratocumulus clouds which closely follow fresh outbreaks of Pc air south of the Great Lakes.

In fresh, unmodified Pc air, icing is practically unknown, except in frontal zones, where it may occur in the air mass being displaced. Icing may also occur in Pc air masses which are overlain by warm air from which rain is falling. This type of icing is due primarily to the presence of the warm, moist air aloft, with the Pc mass furnishing only the low temperature to cause actual freezing.

Convergence within a Pc mass in connection with cyclogenesis or frontogenesis very often produces widespread low clouds. This is not strictly an air mass phenomenon and will not be considered in detail here. (See pp. 311-315.)

After the high winds accompanying the initial appearance of Pc air have diminished, the air mass will be smooth, except in cases where it has had a maritime trajectory. In such instances the instability developed will cause moderate roughness in the lower levels, especially within the stratocumulus or cumulus clouds.

Characteristic Winter Properties of Polar Canadian Air—Source region properties for Pc air are well displayed in the accompanying sounding from Fairbanks. This shows the characteristic intense surface inversion, and thick isothermal layer. Above the isothermal layer appears an upper layer exhibiting a fairly steep lapse rate. Only the portion of the air mass from the surface up to the top of the isothermal layer (2000 meters) is strictly Pc air. The upper layer is the high level polar air found universally at higher elevations in the winter over the Polar region.

The Ellendale data represent only slightly modified Pc air. Here, too, only the lower 2000 meters is Pc air in the strict sense. Various degrees of modification are shown by aerographic material from Royal Center, Boston and Broken Arrow.

In all aerographic soundings the *lift* is the distance that a particle at any level must be lifted in order to saturate it, assuming dry adiabatic cooling.

Summer Properties of Polar Canadian Air Masses—The source properties of Pc air in summer are very different from those prevailing in winter. The terrain in the summer is a typical land surface, which is strongly heated by the sun during the long days of the warm season. This results in moderately unstable surface layers, in contrast to the extremely stable winter conditions. As a result, moisture acquired at the surface is carried aloft, so that the air mass is much more moist throughout than in the winter. Air which moves into central Canada from the Arctic Ocean, or across the mountains from the Pacific Ocean, is ordinarily not distinguishable from that which has remained for a long time over the actual source region. Turbulence and instability causes all of these air mass types to be rather thoroughly mixed, and to have very nearly the same properties in their lower levels during the summer. The upper levels are practically indistinguishable at any time.

The *specific* humidity of Pc air is low compared with other air masses. Its *relative* humidity is generally much lower in summer than in winter since convection during the warm season carries aloft any moisture which may be acquired by evaporation from the surface. Its temperature is low compared with air of tropical origin. Pc air has a very large diurnal temperature variation in the lower layers due to its dryness, which allows rapid radiational cooling at night and insolational heating during the day. Because of the dryness of Pc air at its source, very few clouds are observed in spite of the instability that frequently develops in the afternoon.

Summer Modifications of Polar Canadian Air Masses—The chief modifications of Pc air in the summer, consist in the addition of heat and moisture as it passes southward away from its source. Subsidence effects are practically lacking and the effect of turbulence, as the air passes over mountainous terrain, is unimportant since the inherent instability of the air keeps the lower portions well mixed without outside aid.

The general movement of Polar Canadian air in the summer differs greatly from that in the winter. In cold weather, outbreaks of Pc air move southward as one great uniform mass, showing everywhere remarkably similar properties. In the summer, however, the air flow is more in the nature of isolated streams of air.

AVERAGE Pc_{1-2} AIR IN SUMMER
ELLENDALE, NORTH DAKOTA

Elev. Meters	Temp. °C.	Temp. °F.	D. P. °F.	R. H. %	θ_e °A.	W g/kg.	Lift Meters
Surface (444)	19	66	43	42	312	6.3	1500
1000	16	61	....	45	313	5.6	1500
2000	10	50	....	43	312	3.9	1500
3000	4	39	....	44	314	3.1	1400
4000	−3	27	....	57	318	2.9	800

AVERAGE Pc_{2-3} AIR IN SUMMER
ROYAL CENTER, INDIANA

Elev. Meters	Temp. °C.	Temp. °F.	D. P. °F.	R. H. %	θ_e °A.	W g/kg.	Lift Meters
Surface (225)	17	63	52	68	314	8.3	700
1000	13	55	....	57	311	5.8	900
2000	6	43	....	64	310	4.5	800
3000	2	36	....	43	311	2.6	1400
4000	− 2	28	....	27	314	1.4	1900
5000	− 7	19	....	26	318	1.0	1900
6000	−16	3	....	55	320	1.3	800

AVERAGE Pc_{3-4} AIR IN SUMMER
PENSACOLA, FLORIDA

Elev. Meter	Temp. °C.	Temp. °F.	D. P. °F.	R. H. %	θ_e °A.	W g/kg.	Lift Meters
Surface (4)	23	73	66	79	332	13.4	500
1000	20	68	....	65	330	9.8	900
2000	12	54	....	67	325	7.2	800
3000	7	45	....	57	324	5.0	1000

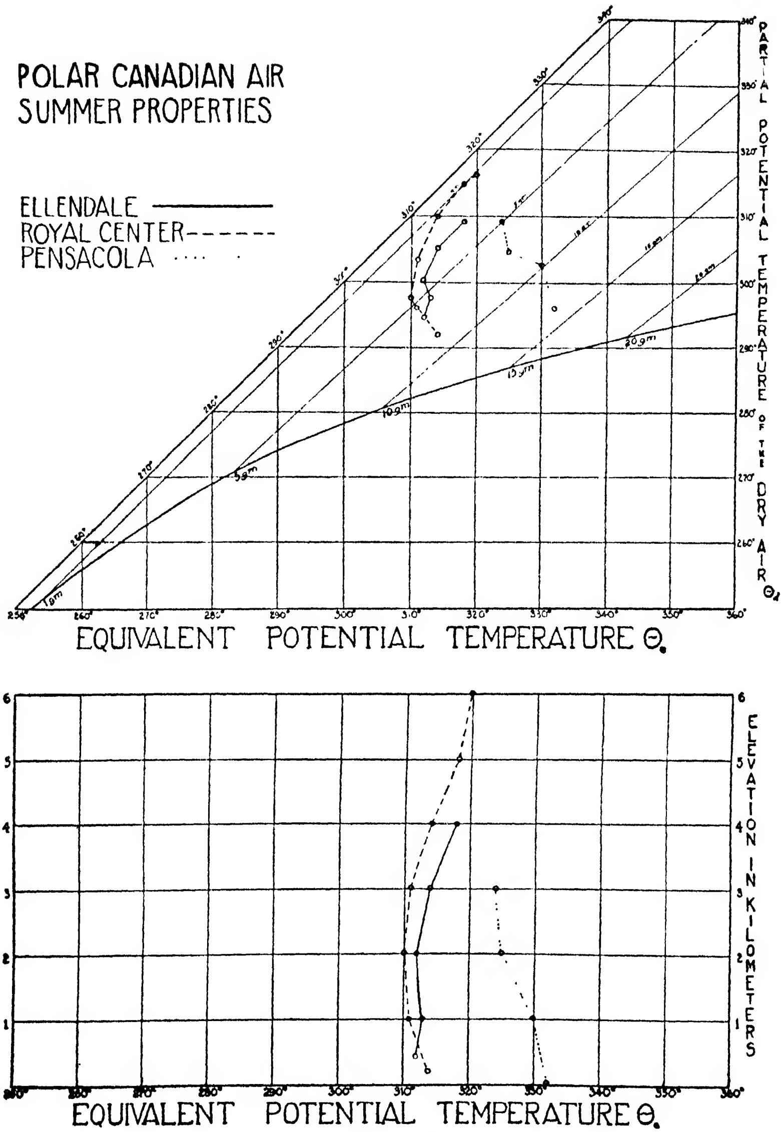

POLAR CANADIAN AIR
SUMMER PROPERTIES
ELLENDALE
ROYAL CENTER
PENSACOLA
PARTIAL POTENTIAL TEMPERATURE OF THE DRY AIR Θ_d
EQUIVALENT POTENTIAL TEMPERATURE Θ_e
ELEVATION IN KILOMETERS
EQUIVALENT POTENTIAL TEMPERATURE Θ_e

FIGURE 45

These arise in localized portions of the source region and they may persist in their direction and type of flow for several days. Air reaching Omaha during such air movements may thus be greatly different from air reaching Cleveland. The air at Omaha may have come directly from Saskatchewan and Mackenzie with practically no trajectory over water bodies, while the air at Cleveland may have passed over the Hudson Bay and the Great Lakes and exhibit greatly different properties. This absence of unified air flow in the summer precludes any comparison between stations which are separated a long distance longitudinally. The Hudson Bay, a vast body of cold water which is frozen for 8–10 months during the year has a very pronounced cooling effect on air passing across it. The Great Lakes, on the other hand, have but little effect on air masses during the summer because of their small area, and the warmth of their surfaces. Pc outbreaks which enter the United States after a trajectory over Hudson Bay are invariably much cooler than those which travel southward through Manitoba, Saskatchewan and Alberta.

As Polar Canadian air moves southward away from its source region during the summer it conforms to Bergeron's cPK classification, to Willett's NPc and to Krick's Pc (occasionally *w*Pc). It is comparatively unstable compared to Pc air in the winter. It becomes increasingly so as it moves southward, and the lower layers acquire heat and moisture from the land over which it passes. Convective instability occurs in the lower 5000–10,000 feet after the air has moved 2 or 3 days away from its source, with the result that mild cumulus activity occurs in these masses soon after they enter the United States. With increasing water content in the intermediate levels, whence it is carried by convection, cumulus activity increases. As a result, air mass thunderstorms may develop during the afternoon within 3 or 4 days after a Pc air mass has left its source region.

Summer Flying Weather in Polar Canadian Pc Air Masses—Ceilings are generally unlimited in comparatively fresh Pc air since the only condensation forms are scattered cumulus clouds. With older Pc masses, convective showers may form, but here too the condensation level is generally rather high, above 2500 feet, so that ample ceilings will be found. Visibilities are generally good during the day because of the dryness, and the turbulence, which carries surface pollution of the atmosphere aloft. At night, because of the high diurnal temperature range and the fairly high humidity,

especially after the air has been away from its source for over 3 days, ground fogs may form for short periods. Turbulence is generally slight above 8000–10,000 feet since this is above the level of convective activity, even during the afternoon. This is not true in old types of Pc air in which thunderstorms have developed. Even in this type of air, turbulence will be noted only in the clouds, for the conditional instability does not cause turbulence until the air has become saturated.

Characteristic Summer Properties of Polar Canadian Air—The source region of Pc air is not represented by any summer soundings. The information from Ellendale, however, serves to give a very satisfactory idea of the source properties. This shows the dryness, relative coolness and slight positive stability of this air.

Modifications in its structure as the Pc air moves southward are shown by soundings from Royal Center and Pensacola. These show increasing temperature and moisture, and a change to convective instability as the air moves southward from its source.

POLAR BASIN AIR MASSES

An air mass type which is found only in the winter season, and then rather infrequently, is produced over the Great Basin region of the western United States when this area is occupied by an intense anticyclone. Under these conditions Polar Pacific air is greatly modified by subsidence effects until it no longer is recognizable as a Polar Maritime type. Krick has named this air *Polar Basin*, after its source region. It is very spasmodic in occurrence, being found frequently some winters, and rarely during others.

Winter Properties of Polar Basin Air—Polar Basin air is much dryer and warmer than Polar Pacific air and very much warmer than Polar Canadian air during the winter. It might be termed *Modified Polar Pacific* air, but the modification has proceeded so far that all of the maritime characteristics have been lost, and a new name seems to be well warranted. When it passes eastward over the Great Plains it brings fine weather, with generally clear skies due to its very low relative humidity and warmth, both of which properties are considerably intensified by foehn effects as the air descends the east slope of the Rocky Mountains. Even when forced aloft, as frequently occurs when it encoun-

Typical Pb Air in Winter Salt Lake City, Utah							
Elev. Meters	Temp. °C.	Temp. °F.	D. P. °F.	R. H. %	θ_e °A.	W g/kg.	Lift Meters
Surface (1280)	− 2	+ 28	24	82	291	3.2	300
2000	+ 5	41	....	39	305	2.6	1600
3000	+ 2	36	....	31	309	1.9	1800
4000	− 4	25	....	25	313	1.2	2000
5000	−10	14	....	23	314	0.7	2100

Typical Pb_2 Air in Winter Detroit, Michigan							
Elev. Meters	Temp. °C.	Temp. °F.	D. P. °F.	R. H. %	θ_e °A.	W g/kg.	Lift Meters
Surface (175)	4	39	32	75	289	3.9	500
1000	8	46	....	40	300	3.1	1500
2000	1	34	....	39	301	2.0	1500
3000	− 5	23	....	48	303	1.7	1300
4000	−11	12	....	74	309	2.0	500
5000	−18	0	....	88	311	1.7	200

Typical Pb_2 Air in Winter Spokane, Washington							
Elev. Meters	Temp. °C.	Temp. °F.	D. P. °F.	R. H. %	θ_e °A.	W g/kg.	Lift Meters
Surface (610)	− 3	27	25	91	282	2.9	100
1000	+ 3	37	....	63	293	2.7	1100
2000	+11	52	....	23	312	2.3	2300
3000	+ 6	43	....	28	317	2.4	1900
4000	+ 2	36	....	22	319	1.5	2300
5000	− 6	21	....	20	319	0.9	2300

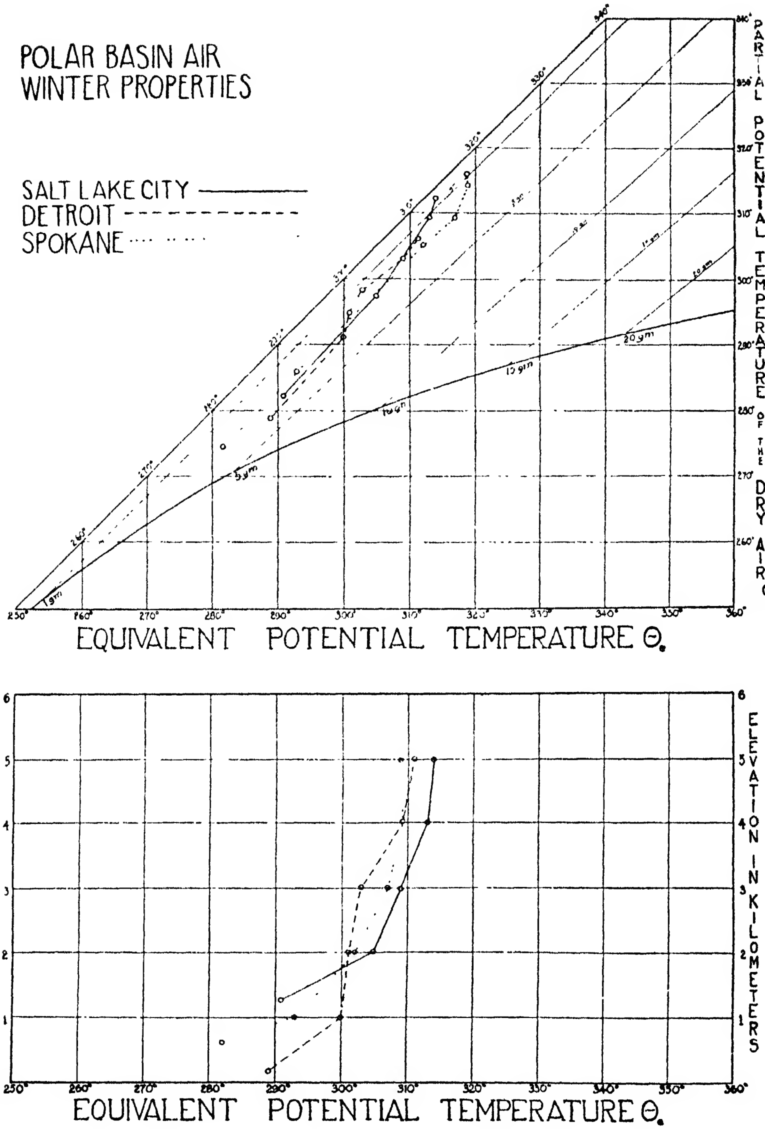

POLAR BASIN AIR
WINTER PROPERTIES
SALT LAKE CITY
DETROIT
SPOKANE
PARTIAL POTENTIAL TEMPERATURE OF THE DRY AIR Θ_d
EQUIVALENT POTENTIAL TEMPERATURE Θ_e
ELEVATION IN KILOMETERS
EQUIVALENT POTENTIAL TEMPERATURE Θ_e

FIGURE 46

ters a wedge of Pc air, its dryness does not allow precipitation or even cloudiness to develop. At all times it is very stable. As it moves well to the eastward, it acquires considerable moisture and pollution in the lower levels. Ground fogs are thus likely to occur after it has moved east of the Mississippi River.

The characteristics of Polar Basin air are best shown by aerographic soundings from Salt Lake City, Utah, which is near the center of the source region. Virtually the same properties are exhibited at Billings, Montana, and Cheyenne, Wyoming, during periods of PB activity. It may be observed at such times that very little fresh Polar Pacific air is added to the Great Basin anticyclone in the lower levels, although it is probable that considerable transport of polar air occurs at high levels in connection with wave action along the tropopause.

Polar Basin air overspreads the entire United States west of the Mississippi River during periods of strong activity and occasionally travels entirely across the country to the Atlantic coast. Often weak outbreaks of Polar Canadian air will occur in the northern and northeastern United States at times when the Great Basin anticyclone is active, causing temporary blocking of the eastward PB flow at the surface. At such times PB air is frequently found aloft over the eastern states and may be confused with Superior air. It is considerably colder than true Superior air, however, due to its northerly origin.

Winter Flying Weather in Polar Basin Air Masses—Flying conditions are generally excellent in Polar Basin air. Clouds and precipitation are almost entirely lacking. Smooth air is characteristic due to its great stability. Fog occurs only after the air has moved well to the east, away from its source region. In short, Polar Basin air produces the finest flying weather of the winter months.

Characteristic Winter Properties of Polar Basin Air—A typical sounding from Salt Lake City shows the source region properties of this air mass. Obvious are the relative warmth and dryness, the strong surface inversion, and the great stability.

The slight modification produced as this air moves away from its source is evident in soundings from Detroit and Spokane, representing approximately the same air that was present in Salt Lake on the day of the sounding mentioned above.

POLAR PACIFIC AIR MASSES

Polar Pacific air dominates the weather of the Pacific coast at least 75% of the time. Very frequently it overspreads much of the remainder of the United States, occasionally completely covering it during strong outbreaks in the winter. At all times it plays an important part in weather disturbances of western and central Canada, all of the United States and most of Mexico, and therefore it deserves considerable attention. Polar Pacific masses which reach the west coast of North America acquire their general properties over interior Alaska and Siberia, in much the same type of continental Arctic or Subarctic environment as Pc masses. The lower levels of these continental Polar masses are then strongly modified in the "secondary" source region of the north Pacific. When they reach North America they are typically maritime in their lower 2 to 4 km.

The paths of Polar Pacific air masses after they leave their source region vary greatly. Occasionally the air takes a direct course from the Aleutian region to the coast of British Columbia and Washington, reaching land in 3 days or even less. At other times it takes a more southerly course requiring 4 or 5 days to reach the continent. On still other occasions it takes a very roundabout southerly path and does not reach the coast of the United States for 6 to 8 days. In the first case the air is relatively cold and dry with perhaps only the lower 2 km modified from the original cP type of air. In the second case the modification may extend through the lower 3 km. In the last case the entire air mass below 4 km will be warmed and moistened. These widely divergent trajectories explain the great variations in the properties of this air as it reaches the Pacific coast. Although the lower levels are thus modified to a varying extent by the ocean surface, the upper levels, about 3 to 5 km, retain their initial continental character, and are in general, indistinguishable from Pc air.

Occurrence of Pp Air in Winter—Outbreaks of Polar Pacific air in the winter are associated with increasing intensity and general southward movement of the Aleutian LOW. This causes air from either Siberia or Alaska to be drawn into the general Polar Pacific source region. At the beginning of these outbreaks the Polar Front is generally well to the north of its average winter position, lying

across the north Pacific Ocean near latitude 60°. The first indications of a fresh outbreak of Pp air are generally:

(1) the presence of well marked negative tendencies in the Aleutian Region with rapid deepening of the general low pressure system, and,

(2) the formation of an active wave along the Polar Front in this area. This initial wave generally moves rapidly southeastward along the front at a velocity varying from 30 to 45 miles per hour. It is first noted along the west coast of North America by general falling tendencies from Cordova, Alaska, northward, followed in 12–18 hours by a widespread rain area.

In most instances this first wave does not cause weather disturbances much to the south of Prince Rupert, B. C. This initial wave is almost invariably succeeded at rather regular intervals by additional waves, with a wave length from crest to crest of approximately 1000 miles. These waves, and the low pressure centers and fronts associated with them, move at relatively very constant velocities at the same time of year, 800–1200 miles per day during the winter and 500–800 miles per day during the summer. The Polar Front often shows a number of waves, fairly equally spaced, extending quite across the Pacific Ocean from northern Asia to North America. Frequently these extend partially or entirely across the United States and Canada. It is very important to keep this fact in mind in making synoptic analyses, since frequently the paucity of weather reports over the ocean may make it necessary to indicate the presence of waves on the basis of only one or two reports. Occasionally the synoptician will find it necessary to draw a complete wave entirely on inference, and will find it necessary to prepare forecasts from this very sketchy analysis!

At times in the winter, the daily velocity of these waves on the Polar Front may be nearly equal to their wave length. They may then appear to be stationary, when viewed on consecutive 24 hour charts, whereas in reality they are moving with a speed of 35 to 50 miles per hour.

It is almost always found, during Pp outbreaks, that each wave extends its effect farther and farther southward. The first one, as mentioned above, generally does not affect the weather much to the

south of northern British Columbia. Succeeding waves then affect a greater and greater area to the south, until toward the end of a series, which generally lasts from 5 to 7 days, the waves will affect the Pacific coast as far south as southern California. This is a fact of great practical utility to the forecaster, for it makes it possible to know in a general way what to expect in the future development of a young cyclone family. It also explains several interesting features of west coast weather: thus, disturbances in the southern California region very frequently occur at intervals of approximately a week, and storms in the northern California, Oregon and Washington area are often preceded by several days of cloudiness. The latter phenomenon is due to the fact that at a given locality such as San Francisco, each of the early waves of a series, which may cause stormy weather to the north, will give rise to only cloudy weather at San Francisco. Thus, each active disturbance in that area is generally preceded by several minor waves, each one of which causes increasing cloudiness. The experienced forecaster will thus not be misled by activity observed along a wave far to the north of this area, if it is an early wave in a series. Toward the end of an intense wave series, the center of the Aleutian LOW may move far to the southward until it lies at latitude 35°–40°. The final waves may affect the Pacific region well into the Trade Wind belt, perhaps causing Kona storm conditions in the Hawaiian Islands and squally weather in the entire Pacific region north of latitude 20°.

When Polar Pacific air approaches the Pacific coast after an extended journey over the ocean, the lower layers are frequently so greatly modified by contact with the warm water they are scarcely distinguishable from Tropical Pacific air. The upper portions of these old Polar masses, however, retain their coldness and convective instability. When they are lifted sufficiently to release it, torrential rains may result. This lifting is generally provided by a warm front type of occlusion (see page 204) in which the old PP mass, retaining its initial instability in the upper levels, rises up over a colder and denser wedge of PP air which is trapped against the coastal mountains.

This type of situation is of the greatest practical importance in Pacific coast forecasting, and is at times very difficult to foresee. The warm front type of occlusion process is very common along the coast of southern Alaska and British Columbia during the colder season and occurs almost invariably during the winter when the PP

air approaches from the west or southwest. At this time of year the land surface in this region is colder than the water surface so that a cold wedge of air lies against the mountains most of the time. To the south, this type of occlusion is less common. The progressive warming of the land surface to the south is more rapid than the water surface, and as a result a close balance frequently exists between the temperature of the air along the coast and the air in the approaching Pp mass. If the Pp mass is *colder* (and denser) than the air along the coast which lies against the coastal mountains, it will displace it, and the weather phenomena following the passage of the front will be only those of air mass type. If the advancing Pp mass is warmer than the air along the coast, however, it will not be able to displace it. It will, instead, pass up over the air along the coast. In doing so, it will be lifted over the coastal mountains and lose much of its moisture. These will frequently be in the form of heavy showers due to the instability of the upper levels of the air mass.

This balance between the temperature of the advancing air and the air along the coast is frequently very delicate. The mere matter of whether the Pp air arrives during the day or night may determine whether the occlusion process shall be of the warm front or the cold front type. Frequently, too, the advancing mass of Pp air may occlude as a warm front type to the north, and as a cold front type to the south. The determination of where it will begin occluding as a cold front is of prime importance to the forecaster, and is a most difficult matter to decide, in view of the utter lack of aerological data in the ocean area to the west.

The importance of knowing which type of occlusion will take place can readily be realized when it is remembered that the cold front type will generally produce a narrow storm area followed by scattered instability showers. The warm front type on the other hand, may produce continued moderate or heavy rain lasting for several days over the coastal area. Once a warm front occlusion is established, it will generally be intensified by the precipitation, which cools the air entrapped between it and the mountains. The fact already mentioned—that the diurnal temperature change may affect the type of occlusion—may be used with advantage in doubtful cases. When a Pp mass of very nearly the same density as the air along the coast passes over the coastal region at night, it may be expected to occlude as a warm front type. In this case the air

along the coast has been cooled more during the night than the air over the ocean. When the same mass passes inland during the day, it may occlude as a cold front type. Now the land surface, and the air above it, has been warmed more than the ocean surface, and the approaching PP mass is denser than the air it is displacing.

As Polar Pacific air passes inland in the winter its potential instability is released by the mountains. Showery weather results and affects the area from the Pacific coast eastward across the various mountain ranges of the western states to the Great Plains. The past history of the PP outbreaks greatly influences their activity as they pass over this rugged region. Air with a short maritime trajectory ($_{3-5}$PP), is invariably extremely unstable due to its coldness, even in the lower layers, and yields heavy rains and snows as it passes inland. These showery conditions often persist for several days, with instability squalls and much cumulus activity. Severe icing conditions for aircraft almost invariably occur in this type of air, particularly in frontal zones. Air masses with a longer maritime trajectory ($_{5-8}$PP) do not show nearly as much activity, since the lower and intermediate levels have become fairly well stabilized. The passage of the cold fronts of these air masses is accompanied by only light or moderate showers and the instability conditions within the air mass last but a day or so. Icing is present in the immediate vicinity of the cold front and to a slight extent in the cumulus clouds within the air mass, but is much less severe than in the fresher outbreaks.

As the Polar Pacific masses move eastward across the Rocky Mountains they may behave in two very distinct ways, depending on the synoptic situation in the Plains region. If the area to the east of the mountains is occupied by a relatively warm type of air (RPP, RPC or TG), the PP air will descend the east slopes of the Rockies as a rather cool and dry current, warmed to some extent by foehn effects, and dessicated from its passage across the mountains. The forward edge of the PP current is generally an old occlusion of the cold front type, extending for several thousand miles more or less parallel to the mountains, in a weak trough. Frequently this occlusion may be traced the entire distance, from the southern Rocky Mountains region northward through the United States, Canada and across Alaska to the center of the Aleutian LOW.

As the PP air moves down the eastward slope of the mountains it is generally so dry that it gives rise to little precipitation itself.

However, the circulation induced by the trough of low pressure along the Pp occlusion, frequently draws warm air up from the south, with the formation of a warm front and a secondary cyclone. As this new system deepens and moves east, or northeast, it may give rise to a fairly large precipitation area if the air that is being drawn in from the south is moist. This latter point is important in the synoptic analysis. If the air drawn into the cyclone is returning Pp or Pc with little or no maritime trajectory, the new cyclone may deepen very little and cause but little precipitation. If, however, the air moving northward is of Tg origin, or a returning Pp or Pc current with several days trajectory over the Gulf of Mexico, the system may deepen rapidly and cause widespread storm conditions.

If the Plains region, on the other hand, is occupied by a Pc mass that is fairly fresh and cold, the Pp air, because of its lower density will move eastward *aloft* over the dense wedge of Pc air at the surface. Under these conditions a very widespread and persistent blizzard may develop in the area east of the Rocky Mountains, caused by the precipitation falling from the over-running Pp air. The upper boundary of the Pc air mass is marked by a very strong inversion, with the result that surface pollution becomes concentrated in the air near the surface with consequent very poor visibility. This is often accompanied by strong surface winds and continual snow flurries. The passage eastward of the upper cold front, which makes the advancing edge of the Pp air mass, is marked at the surface by a weak trough and a line of marked tendency discontinuity (see page 206). Occasionally this surface trough may give rise to very weak cyclonic circulation *within* the Pc mass at the surface, but this surface circulation has nothing to do with the actual weather phenomena.

The Pp air descending the east slope of the Rockies, may at other times advance some distance eastward across the Mississippi Valley region at the surface, before it encounters air sufficiently cold to force it aloft. It will then occlude as a warm front type occlusion and again move eastward aloft as described in the preceding paragraph. This process is described in some detail on pages 201-208. Aircraft icing is usually rather severe in the conditions just described. Much water in liquid form is present in the Pp air, and the marked turbulence near the front gives rise to large droplets, both conditions being conducive to icing. Furthermore, under these conditions planes must take off in the very cold Pc air

at the surface, and any moisture that they encounter as they enter the over-running Pp current will freeze instantly on impact.

Although Pp air loses much of its moisture in passing over the mountain ranges of the western United States, nevertheless it retains sufficient humidity to make it readily distinguishable from Pc air during the colder months, as it moves eastward across the United States and Canada. It is generally much warmer than the continental air during this time of year, due to the fact that it was initially warmer in the lower levels and because it is subjected to considerable dynamic heating in descending the east slope of the Rocky Mountains. The upper levels, above 3–4 km, even in the winter, however, are very similar to the upper levels of Pc air—dry and cold. After Polar Pacific and Polar Canadian air masses have remained over the surface of the central and eastern United States for 3–5 days, they generally have become so modified by surface heating, loss or gain of moisture, and intermixing, that even in the lower levels, they are indistinguishable, and act in exactly the same manner in cyclones in which they may be involved.

Winter Flying Conditions in Polar Pacific Air—The instability showers that follow fresh outbreaks of Pp air on the Pacific coast and in the Great Basin cause poor flying weather for 1–3 days after the air arrives in this region. This is particularly true in the mountainous regions, where severe squalls often persist for several days. Clouds are of the cumulus type, and ceilings and visibilities are generally ample in the coastal and inter-mountain valleys except during actual precipitation, when scud clouds may cause very low ceilings and visibilities. Within 24–48 hours after appearing over the Great Basin region the Pp air begins to become stabilized. It often returns toward the coast as a warm, dry foehn current.

The air in fresh outbreaks of Pp air is generally rather rough due to its instability. This soon disappears after subsidence commences. Icing is pronounced in fresh outbreaks, particularly near frontal zones, and over the higher mountains. Generally within 8–12 hours following frontal passages icing is rarely found except in the heavier squalls over the high mountains.

After Pp air passes eastward over the Rocky Mountains, it acts very similarly to Pc air if it descends to the surface. If it is forced aloft by cold Polar Canadian air at the surface, it may cause a wide zone of blizzard conditions with very poor flying weather. Generally it is somewhat more unstable than Pc air, so that visibilities are

Typical PP_1 Air in Winter
Fairbanks, Alaska 1-6-37

Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (150)	−14	7	6	96	262	1.3	0
1000	0	32	....	40	287	1.7	1400
2000	− 3	27	....	34	292	1.3	1500
3000	− 8	18	....	45	298	1.4	1200
4000	−10	14	....	80	310	2.1	500
5000	−15	5	....	85	314	1.9	300

Average ${}_{2\text{-}3}PP$ Air in Winter
Seattle, Washington

Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (5)	8	46	36	66	292	4.4	700
1000	0	32	....	64	289	2.7	700
2000	− 8	18	....	64	288	1.5	900
3000	−14	7	....	52	289	0.8	1200
4000	−19	−2	....	35	294	0.4	1600

Average ${}_{3\text{-}4}PP$ Air in Winter
San Diego, California

Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (5)	14	57	46	63	301	6.1	900
1000	8	46	....	74	302	5.3	500
2000	1	34	....	57	300	2.8	1000
3000	− 6	21	....	40	299	1.3	1400

Average ${}_2PP_2$ Air in Winter
Ellendale, North Dakota

Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (444)	− 1	30	26	83	284	3.0	300
1000	7	45	....	43	299	3.0	1400
2000	+ 1	34	....	44	300	2.2	1400
3000	− 7	19	....	48	301	1.5	1200
4000	−14	7	....	60	302	1.1	900

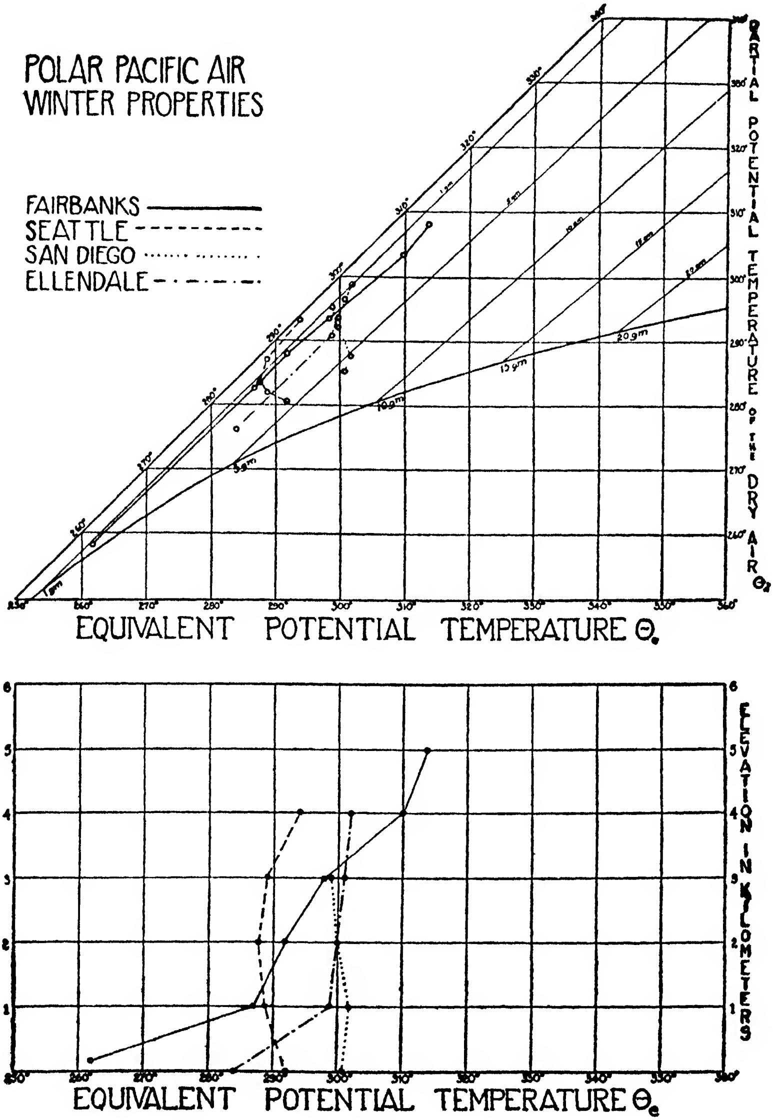

POLAR PACIFIC AIR
WINTER PROPERTIES
FAIRBANKS
SEATTLE
SAN DIEGO
ELLENDALE
PARTIAL POTENTIAL TEMPERATURE OF THE DRY AIR Θ_d
EQUIVALENT POTENTIAL TEMPERATURE Θ_e
ELEVATION IN KILOMETERS
EQUIVALENT POTENTIAL TEMPERATURE Θ_e

FIGURE 47

better. Turbulence, except that due to mechanical effects, is almost entirely lacking after the air passes over the mountainous region.

Characteristic Winter Properties of Polar Pacific Air—No aerographic soundings are available in the Pp source region. Fairbanks, however, probably represents fairly reliably the source region properties of this air mass. The water vapor content and the temperature are undoubtedly somewhat too low, but probably not materially. Comparatively little reliable data is available from Fairbanks concerning Pp air masses, since most Pp invasions of Alaska become well mixed with Pc air. Furthermore, the Fairbanks aerographic station has been in operation but a short time.

A typical sounding in fresh Pp air with one day's continental trajectory appears in figure 47. This shows more stability than is probably present in the source region. The surface temperatures are also colder than might be expected there. It is probable that the equivalent potential temperature is approximately constant, or shows only a slight increase aloft, in the source region. The surface temperatures there cannot be far from freezing.

Insufficient study of Pp air masses has been carried on to allow their separation into continental and maritime types. Undoubtedly, much variation occurs in the properties of Pp air. Some of it crosses the Pacific Ocean immediately after a long sojourn over Siberia, while some proceeds around the periphery of the Pacific HIGH with a very extended maritime trajectory.

Polar Pacific air masses which reach Seattle may have widely different past histories. The aerographic data at hand represent an average of the various types. They show characteristic convective instability through most of the lower 3000 meters, low moisture content aloft, a steep lapse rate throughout.

A considerable increase in water vapor content throughout the lower 3000 meters is noted as Pp air moves southward. This is undoubtedly due to turbulence caused by the high wind velocities along the rough coast line. San Diego soundings show this clearly, as well as the continuation of convective instability throughout most of the air mass. The temperature also increases throughout due to the same turbulence effects that cause the moisture to increase.

The Ellendale soundings indicate the continental influence plainly. The air becomes stabilized, cooled in its lower levels, and dessicated, from passage eastward over the mountains.

Summer Properties of Polar Pacific Air—Polar Pacific air in the summer as in the winter has its true source in the continental regions of central Alaska and eastern Siberia. The modifications over the secondary source region of the north Pacific, however, are greatly different from those during the winter. During the warm season of the year, the ocean is uniformly colder than the adjacent land surfaces and therefore tends to *stabilize* any air passing over it from the continents. Because little or no instability is developed in the air during the summer, there is little modification of any but the surface layers, as the air moves across the ocean. Thus, when Polar Pacific air moves inland during the summer it is practically indistinguishable from Polar Canadian air except by its trajectory, and both types of air behave in essentially the same manner.

Because of the characteristic summer stability of Pp air, the convective showers that accompanied it in the winter are almost wholly absent, except occasionally in the higher mountains along the west coast. The movement of Pp masses southward from the source region is considerably slower than in the winter because of the relatively sluggish air flow during the summer season. The paths are somewhat different, too, since the Pacific High Pressure Area dominates the circulation of the north Pacific as far north as latitude 40°–50°, and as a result outbreaks of Pp air never take the southerly path that they occasionally do during the winter, but on the contrary, travel rather directly from the Aleutian region to the Pacific coast around the periphery of the Pacific HIGH. Frequently, too, a general southward air movement occurs from the Alaskan and Mackenzie regions in which Pp air on the west moves in the same general current that brings Pc air farther east, emphasizing the close relationship of these two air masses in the summer.

The approach of a Pp mass to the Pacific coast region in the summer causes an increased onshore gradient that frequently gives rise to prefrontal stratus or stratocumulus clouds. These cloud decks, which are especially characteristic of the summer weather of the Pacific coast region, are discussed in some detail in chapter 15. It will suffice to say here that in spite of the thermodynamic stability of Pp air in the summer a rather well marked turbulent layer is generally present in the lower 2000–3000 feet. This is a result of turbulence due to the fairly high wind velocities in the Pp masses and to the ruggedness of the topography along the Pacific

Average $_{2-3}$Pp Air in Summer Seattle, Washington							
Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (5)	17	63	51	62	308	7.1	1000
1000	9	48	....	91	308	6.8	400
2000	5	41	....	60	307	3.9	900
3000	1	34	....	42	309	2.3	1400
3500	− 2	28	....	33	310	1.7	1600

Typical $_{3-4}$Pp Air in Summer San Diego, California							
Elev. Meters	Temp. °C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (5)	18	64	58	79	319	10.3	400
600	13	55	....	95	317	9.5	100
1000	22	72	....	40	328	7.7	1700
2000	21	70	....	20	324	3.5	3000
3000	14	57	....	27	328	3.6	2300
4000	6	43	....	36	330	3.4	1600
5000	− 2	28	....	40	330	3.0	1000

Typical $_{3}$Pp$_{2}$ Air in Summer Salt Lake City, Utah							
Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (1280)	16	61	43	53	322	7.1	1100
2000	21	70	....	33	334	6.7	2000
3000	14	57	....	34	331	4.6	1900
4000	6	43	....	43	332	4.1	1400
5000	− 2	28	....	55	331	3.2	1000

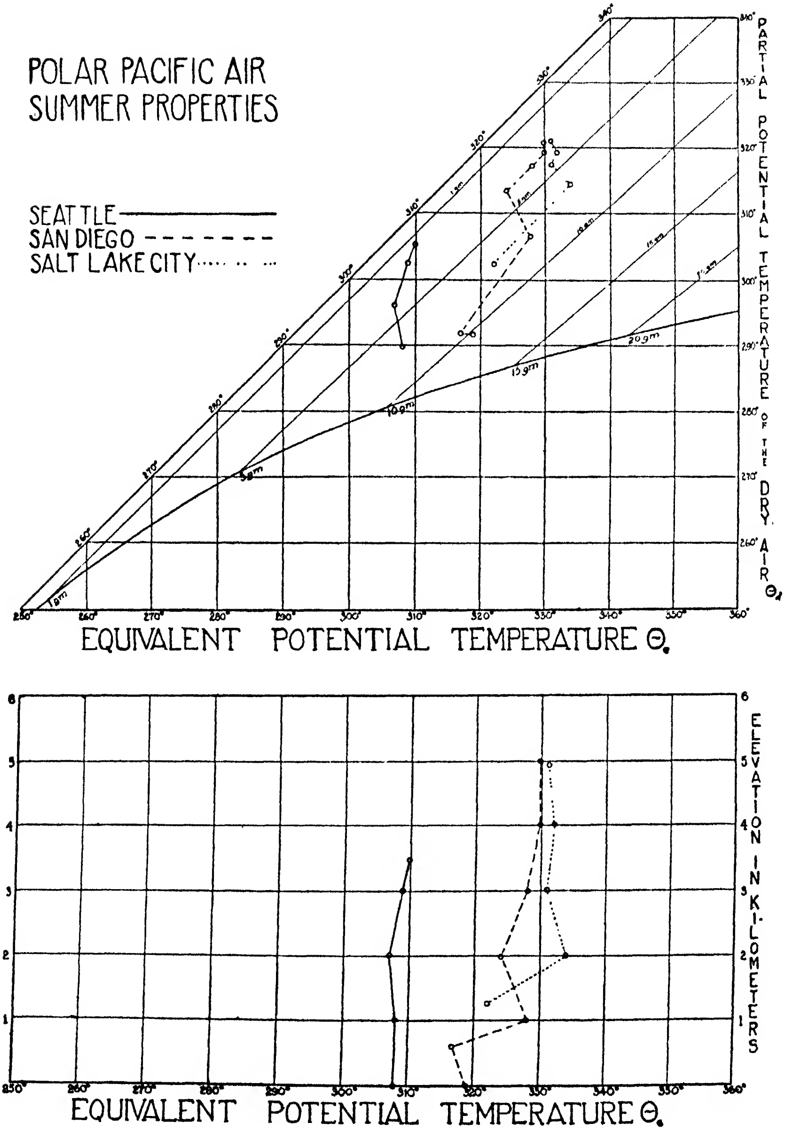

POLAR PACIFIC AIR
SUMMER PROPERTIES
SEATTLE
SAN DIEGO
SALT LAKE CITY
EQUIVALENT POTENTIAL TEMPERATURE Θ_e
PARTIAL POTENTIAL TEMPERATURE OF THE DRY AIR Θ_d
EQUIVALENT POTENTIAL TEMPERATURE Θ_e
ELEVATION IN KILOMETERS

FIGURE 48

coast. It is important to remember also that this turbulence keeps the lower layers of the air well mixed. As a result, a fairly steep lapse rate is produced and the moisture acquired from contact with the water surface is carried aloft. This moisture is not carried much above the 3000 feet level, and above this the air is as dry as Pc, or even dryer, due to subsidence effects in passing around the Pacific HIGH.

Since Polar Pacific air in the summer is indistinguishable from Polar Canadian after it has passed eastward over the mountains it will not be considered separately and, for a discussion of its summer properties in this region, the reader is referred to the section on summer properties of Polar Canadian air.

Summer Flying Weather in Polar Pacific Air Masses—Flying conditions are generally excellent in fresh Pp air with clear skies or scattered cumuli over the mountains and little risk of fog or low stratus. In modified Pp air along the coast, especially in regions of cold surface water temperatures, low stratus clouds or ground fog are frequent. Except in regions of unusual surface heating, Pp air is smooth. Visibilities are generally excellent except in old modified masses along the coast. East of the Rocky Mountains, Pp air exhibits the same flying weather as Pc air (see page 122).

Characteristic Summer Properties of Polar Pacific Air—No data at all are available to indicate the source region properties of Pp air during the warm reason. Soundings from Seattle, however, probably give a fairly close approximation to these properties.

As Pp air moves southward along the Pacific coast it becomes warmer and more moist. The most important change is the formation of a surface turbulent layer with a remarkably steep lapse rate, approximating the dry adiabatic. This layer even appears at Seattle, but is much more apparent at San Diego. Above this surface layer a sharp inversion appears, above which the air is very dry and stable. This turbulence inversion marks the top of the stratus fog which is so characteristic of summer Pacific coast weather.

After Polar Pacific air passes inland during the summer it becomes, within a very short time, indistinguishable from Polar Canadian air. The Salt Lake sounding is typical of summer conditions in Pp air. The higher levels are almost identical with the air as it reaches the coast, and the inherent stability of the air continues. Farther to the east Pp air behaves in all respects like Pc air and soundings in the two air masses are practically identical.

POLAR ATLANTIC AIR MASSES

When Polar Canadian air moves over the waters of the north Atlantic Ocean off the Newfoundland and New England coasts, it is considerably modified in its lower levels. This region is one of uniformly cold surface water for that latitude, 35°–40° F. in winter and 50°–60° F. in the summer. During the late spring and summer the waters of this region are much cooler than the adjoining continent, and it is at this time that Polar Atlantic air masses are of the greatest importance along the Atlantic coast region. Their source is a typical "secondary" source region in which Pc air is modified to a maritime type in its lower levels. Polar Atlantic air does not have nearly the importance of Polar Pacific air in controlling North American weather since the normal west to east air flow in this latitude does not allow it to reach the continent except rather infrequently, and then only to affect the Atlantic coastal area. Air of Polar Atlantic characteristics is very important over the North Atlantic Ocean and northwestern Europe and has been briefly described by Schinze. Its characteristics in the European region are outlined in chapter 8.

Winter Properties of Polar Atlantic Air—Since the usual direction of movement of air masses across the source region of PA air is away from the continent, it is only occasionally that this air reaches the coast. It does so under two general situations:

1. A Pc anticyclone may move offshore between Newfoundland and the British Isles. Air from its southwestern limb may pass over the cold waters of the North Atlantic, reaching the east coast of North America as a NE to ESE current. Rarely does PA air in this situation affect any but the immediate coastal region.
2. A deep LOW in the upper Ohio River Valley, or on the Atlantic coast between Cape Hatteras and Boston will frequently bring about the influx of PA air as a NE to E current. Under this situation the PA air may travel a considerable distance westward and appear as far inland as the Great Lakes area and the upper Mississippi Valley.

Polar Atlantic air is much more stable than Polar Pacific air because of its comparatively brief sojourn over relatively cold waters

and because the returning Pc mass from which it is generally subsiding and becoming stabilized. This subsidence causes the PA air to be relatively shallow and to be overlain by warm dry air. Generally a rather steep lapse rate prevails in the lower 2000-3000 feet due to turbulent mixing and surface heating.

The weather accompanying an invasion of the coastal region by PA air is very characteristic—low stratocumulus clouds, from which light misting rain or light snow flurries are falling. The lack of convective instability inhibits the formation of instability showers, which are so characteristic of Polar Pacific outbreaks along the Pacific coast. The rapid change in weather which occurs when PA air appears along the New England coast is at times very striking. Polar Canadian air in the region is usually accompanied by clear skies, low temperatures and north winds. These are frequently followed within the course of 1-3 hours by a low cloud deck with light rain or snow, a temperature increase of 15°–20°F., and a shift in wind direction to the northeast. This is the typical "northeaster" of the New England coast. Frequently the onset of cyclogenesis along an old front which has passed offshore will result in the formation of a wave off the southern New England coast. This often gives rise to high wind velocities in connection with these outbreaks of Polar Atlantic air.

Winter Flying Conditions in Polar Atlantic Air—In general, flying weather is uniformly poor in this air mass. Ceilings are low under the stratocumulus clouds, visibilities are poor in the misting rain or snow, icing conditions are pronounced due to the turbulent instability of the lower levels and the maritime history of the air, surface winds are often strong and gusty. When PA air passes westward over the Appalachian Mountains, most of these properties are greatly lessened in severity by the passage over the mountains, so that PA air in the upper Ohio or Mississippi Valleys is indistinguishable from Pc air in its effect on flying.

Summer Properties of Polar Atlatic Air—PA air occurs most frequently during the spring and summer, for then the ocean surface in its source region exhibits the most pronounced subnormal temperatures. At these times, a quasi-stationary HIGH often forms over the source region off Newfoundland and Nova Scotia after a Pc outbreak has moved out to sea. This circulation feeds PA air into the New England region and, under favorable pressure distribution, well down the Atlantic coast, occasionally as far south as northern

Florida. The characteristics of PA outbreaks in this season of the year are marked coldness with respect to the warm continent (mPK), a rather thin layer of truly maritime air (2000–4000 feet), generally high wind velocities, a marked inversion at 2000–4000 feet with a stratocumulus cloud deck below it. Precipitation does not occur except in connection with cyclonic activity. Convergence in the cold mass may then cause a light misting rain, or overrunning warm air may result in a widespread rain area.

The rather steep lapse rate which PA air exhibits as it reaches the continent is due to turbulence, since the cold water surface over which it passes precludes any possibility of thermal instability. The specific humidity differs but little from Pc air at this time of year. The relative humidity is much higher, however, because of the marked cooling over the cold source region. Apparently the only modifications of the original Pc properties consist in surface cooling and the establishment of a turbulent layer 2000–4000 feet thick, with consequent low temperatures and high relative humidities in the lower levels.

The appearance of PA air in the summer is marked by a rapid drop in temperature (15°–25° F.) as the warm continental air is displaced. A stratocumulus cloud deck generally appears with the maritime air, but precipitation may be expected only in regions of considerable frontal activity, where convergence or active overrunning are present.

Summer Flying Conditions in Polar Atlantic Air—The stratocumulus clouds associated with PA air in the summer generally are accompanied by ceilings of from 500–2000 feet and do not in most cases constitute a hazard to airline flying. Only when convergence or overrunning by warm moist air occurs do the ceilings drop below about 500 feet. In such cases the misting rain and low clouds give rise to generally poor flying conditions, with both ceiling and visibility becoming very low. It is thus highly important in connection with PA outbreaks to determine the possibility of wave formation along fronts within the forecast area.

In the New England region, wave formation just offshore may cause the very sudden appearance of misting rain, with practically zero ceilings and visibilities over a wide area along the coast. It is thus very important to study the synoptic situation over the Atlantic Ocean east of New England with great care during incursions of PA air, in order to detect the first signs of wave formation. This

is an important matter at any time, but vitally important with PA air, because of its high relative humidity, and the consequent speed with which convergent activity may affect a widespread area.

Characteristic Properties of Polar Atlantic Air—Since PA air has essentially the same properties as Polar Canadian air, except in its lowest levels, characteristic curves for it are not presented here.

TROPICAL GULF AND TROPICAL ATLANTIC AIR MASSES

These air masses are of great importance to the weather of central and eastern United States. They originate in the Gulf of Mexico and Caribbean Sea area, and in the general high pressure belt of the north Atlantic Ocean between latitudes 15° and 30°. The surface temperatures of this "primary" source region are uniformly high, ranging from 70° F. at the coldest near the coast of the southeastern United States, to slightly over 80° F. in the Caribbean Sea. The surface properties throughout this vast area are so nearly uniform that no appreciable difference in air masses originating at various points can be observed. The only differences are due to the varying trajectories which they take in passing over the source region.

Tropical Gulf air masses are drawn into the United States region in winter, under the influence of well developed low pressure areas in the central or lower Mississippi Valley, or over the southeastern states. In summer the general monsoon circulation of that season, between the Gulf of Mexico and the southern United States, brings them over the continent. These bodies of air are drawn from the northern portion of the trade wind belt, through the Caribbean Sea and Gulf of Mexico, and thus are of true equatorial origin. In fact, a study of aerographic soundings from Miami shows that TG air in the summer is practically indistinguishable from air over Batavia, Java, which lies at 6° S. (cf. figs. 50 and 56). This is not surprising, of course, since the source regions in both instances are very similar—open ocean surfaces with nearly equal temperatures. Both are situated on the border of one of the subtropical belts of high pressure, Miami on the northern border of the northern belt and Batavia on the northern border of the southern belt. Since Batavia is very close to the equator, subsidence

effects are not as pronounced as at Miami where they occasionally give rise to warm and dry strata at high levels.

Tropical Atlantic air generally appears in the eastern United States as a south or southeast current, resulting from unusual development of the Bermuda HIGH, following a particularly strong outbreak of Polar air over the Atlantic Ocean. This indicates clearly that TA air is simply Polar air which has been modified in the course of its travel around the periphery of the Bermuda HIGH. During this journey it acquires tropical maritime characteristics. The remarkable fact is that this air, with an oceanic trajectory of only 5–7 days is practically indistinguishable from true Equatorial air. Tropical Gulf air, for instance, has in most cases remained over tropical waters for several weeks. This illustrates the important fact that an air mass over a water surface very rapidly acquires a vertical temperature distribution which represents equilibrium conditions with respect to the underlying surface. That this temperature distribution aloft is the result of radiation, rather than convection effects, is attested by the rapid decrease in moisture aloft, a condition which would certainly not be found with active convective mixing. Only in cases of very rapid movement of Polar air across a warm water surface can it maintain convective instability—as soon as the air flow decreases in velocity, it rapidly becomes stabilized. The rapid modification produced by water surfaces probably is due to both their mobility and to the high specific heat of water.

It has been pointed out that Tropical Gulf and Tropical Atlantic air masses are practically indistinguishable, even at high levels, so any distinction between them is meaningless from a practical point of view. Since Tropical Gulf air is the more common type in the United States, that designation will be used interchangeably for both air masses. The only distinction possible may be made on the basis of their trajectories. Even this is difficult along the eastern seaboard of the United States, when air from the Gulf of Mexico and the Bermuda region may both be involved in widespread low pressure systems.

An important distinction must be made, however, between true Tropical Gulf air, and Polar masses which pass briefly over the waters of the Atlantic or the Gulf of Mexico and then return over the United States as southerly currents. The surface properties of the returning Polar air may approximate those of the TG air, but

aerographic ascents invariably show that the upper levels have not arrived at equilibrium with the water surface. The behavior of this slightly modified Polar air is greatly different from true Tg air. This emphasizes the necessity of observing air masses after they leave the borders of the continent since their trajectories over the adjacent waters influence profoundly their properties when they return. In general it may be stated that approximately 5–7 days are required for a Polar mass to reach equilibrium with a warm water surface. During this period its properties undergo a gradual transformation from their original values to those of a typical Tg air mass. If the air returns after a 2 or 3 day maritime trajectory, it will be considerably modified, of course, but it will not show the characteristic convective instability of Tg air. This is especially important in summer, for convective showers are rarely found in such returning Polar masses, whereas they occur almost inevitably in Tg air. The failure to distinguish properly between these returning Polar masses and true Tg air thus leads not infrequently to erroneous forecasts.

Winter Properties of Tropical Gulf Air—As Tropical Gulf air masses in the winter move northward from their source region over the southern United States they are modified very slowly. They reach the central and northern Mississippi Valley and eastern seaboard states, with little change other than a slight decrease in temperature throughout. The lower levels of this air generally are nearly saturated. As a result even the slight lift imparted by the gently sloping Mississippi Valley is sufficient to cause the formation of widespread stratus and stratocumulus cloud decks. These still further inhibit alteration of the air as it moves northward.

These slow changes may be greatly accelerated if the ground is snow-covered. Melting of the snow may then cause intense cooling, but even so the changes are comparatively much slower than with southward moving Pc currents. The rapid modifications in the Polar Canadian air mass properties in such cases very frequently result in the formation of secondary fronts. In situations where Tg air moves northward over a snow-covered surface it is rapidly changed to an exceedingly stable air mass in the lower levels, with the formation of fog and mist and resulting very poor visibilities. Similarly, when Ta air moves northward over the cold water along the New England coast and over the Maritime Provinces and Newfoundland, it generally develops deep and persistent fogs.

Because of the relative dryness of the upper levels of Tg air in the winter and the absence of the extreme surface heating that is present in the summer, instability showers are rare in the cold season. Convectional instability is still present, however, as may be seen from an inspection of figure 49. If this is released by sufficient lifting along a frontal surface, heavy showers or snows may develop. This is particularly important in the Appalachian region where topographic lifting is added to any frontal lift available. Many of the severe winter floods in this region are attributable to overrunning by Tg and Ta air, above warm fronts which lie over the mountains. The unusually severe floods of the early spring of 1936 were caused by this situation. They were rendered especially destructive by melting of the deep snow cover by heavy rains falling from the Tg air. Since rains of this type are likely to be very widespread, because of the uniformity of the air and the gentle slope of the warm front surface, they may affect the entire drainage basins of major rivers.

Occasionally, in the late fall and early winter, the instability in Tg masses may be sufficient to produce thunderstorms as far north as the northern border of the United States, although these situations are rather uncommon. In the late winter months thunderstorms are rarely encountered north of latitude 35°, since modification of the Tg mass is generally sufficient to destroy much of its original convective instability. Temperatures are generally so low that rain drops are not present in sufficient quantities to produce the large electrical charges, resulting from their shattering, necessary for thunderstorm formation. Occasionally winter thunderstorms are produced from the shattering of snow flakes, caused by their collision, but this process is relatively unimportant.

The change in characteristic convectional instability, as the Tg masses proceed far from their source, is well illustrated by a comparison of the Tg curves in figure 49. Of particular interest is the rather unusual stability exhibited by the higher levels of Tg air at Boston. This is largely a result of the high values of specific humidity at these levels. Willett explains this by the trajectory taken by this type of air in arriving at the New England region. This air, of relatively recent Polar origin, has passed over the northern portion of the Tropical maritime source region. During this journey it has acquired considerable moisture yet has been little affected by subsidence. It reaches New England, after passing across

Typical Tg Air in Winter
Miami, Florida

Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (3)	25	77	71	82	343	16.3	400
1000	20	68	....	82	339	13.3	400
2000	13	55	....	83	333	9.8	400
3000	8	46	....	66	329	6.2	600
4000	3	37	....	67	332	5.2	600
5000	−4	25	....	37	326	2.1	1400

Average Tg_{1-2} Air in Winter
Broken Arrow, Oklahoma

Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (233)	20	68	61	79	328	11.8	500
1000	14	57	....	84	324	9.7	300
2000	12	54	....	42	318	4.8	1400
3000	7	45	....	35	318	3.1	1700

Average Tg_{2-3} Air in Winter
Royal Center, Indiana

Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (225)	18	64	60	89	324	11.3	300
1000	13	55	....	96	322	9.6	100
2000	8	46	....	54	314	4.5	1100

Average $_{2-3}$Ta Air in Winter
Boston, Massachusetts

Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (4)	14	57	54	88	310	8.8	200
1000	14	57	....	59	314	6.5	900
2000	9	48	....	70	319	6.2	600
3000	2	36	....	75	318	4.6	500
4000	−4	25	....	65	319	2.9	700

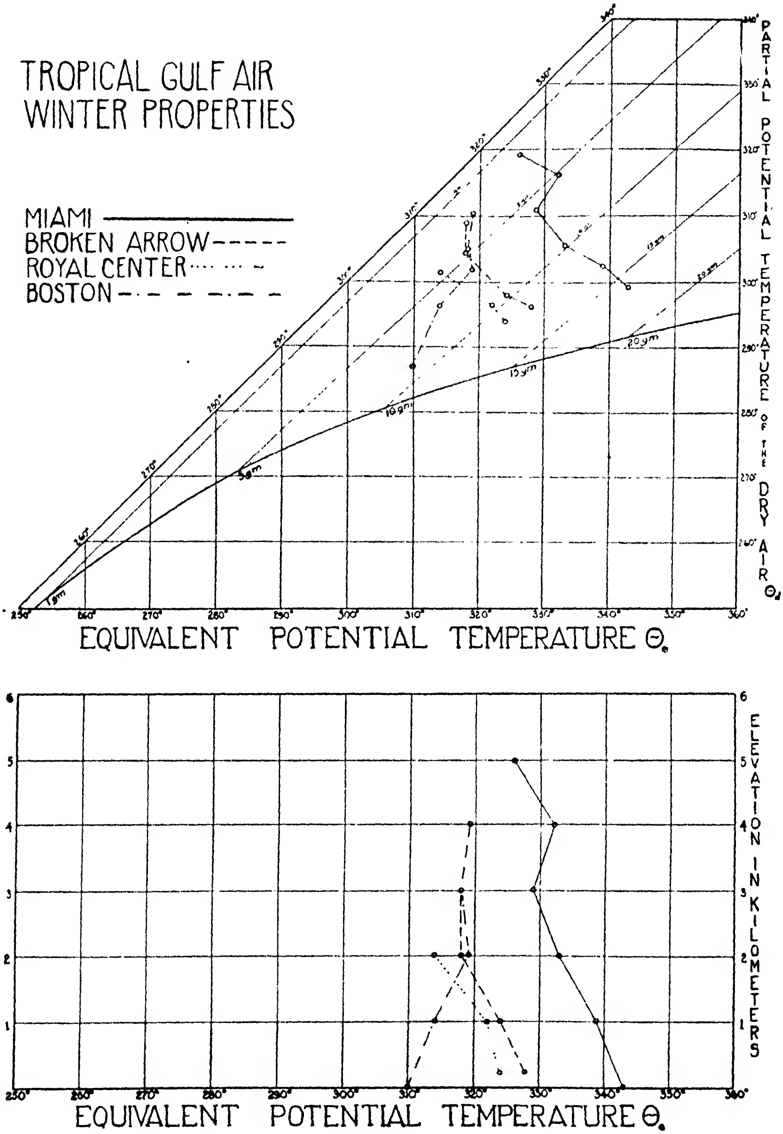

TROPICAL GULF AIR
WINTER PROPERTIES
MIAMI
BROKEN ARROW
ROYAL CENTER
BOSTON
PARTIAL POTENTIAL TEMPERATURE OF THE DRY AIR Θ_d
EQUIVALENT POTENTIAL TEMPERATURE Θ_e
ELEVATION IN KILOMETERS
EQUIVALENT POTENTIAL TEMPERATURE Θ_e

FIGURE 49

the cold water off the coast, as a deep current, stable throughout, with relatively high values of specific humidity even at upper levels. It is important to note the rapid modification which a cold water surface produces on warm tropical air passing across it in contrast to the slight effect of a land surface on the same type of air. As before, the mobility and high specific heat of water are responsible.

Winter Flying Weather in Tropical Gulf or Tropical Atlantic Air Masses—Fresh Tm air near its source is characterized by variable, thin decks of stratocumulus clouds. As it moves northward over cold surfaces and as it is lifted by the land surface, the clouds thicken and lower, often with mist or drizzle and low fog, especially at night. During the day the cloud decks generally dissipate, or at least become broken with ample ceilings. Visibilities are usually good, unless the lower levels become highly stabilized by passing over snow-covered surfaces. Then they may be poor due to smoke pollution. Icing is seldom, if ever, encountered as a pure air mass phenomenon in Tm air. It may be rather serious, however, when Tm air is lifted along active fronts (see page 335). Smooth air is the rule within Tm air masses, due to the stable nature of the air as found over the continent. After the air is lifted along frontal surfaces or over mountainous terrain, however, it usually becomes turbulent due to the release of its convective instability.

Characteristic Winter Properties of Tropical Gulf Air—Miami is close enough to the Tg source region to represent the source properties satisfactorily. Modification as the air passes northward over the land are indicated by soundings from Broken Arrow and Royal Center. These show the increasing stability as the air moves over the cold land surface. The very pronounced stabilizing effect of the cold water of the north Atlantic is shown by the Boston sounding.

Summer Properties of Tropical Gulf Air—The summer weather of the central and eastern United States is controlled chiefly by Tg air. Even the southwestern states occasionally come under the influence of this important air mass. This is largely due to the summer monsoon effect between the strongly heated continent and the relatively cooler waters adjacent to it. When increasing southeasterly circulation, caused by the general intensification of the Bermuda HIGH in the summer, is added to the monsoon flow from the Gulf region, it is little wonder that Tg air is of great importance to so large a portion of North America. Polar out-

breaks are comparatively weak during this season and the Polar Front generally lies north of the Great Lakes.

The source properties of Tg air in the summer are in general very similar to those in the winter, except that the temperatures over the open ocean are somewhat warmer, and along the Gulf coast, 15°–20° F., warmer. As a result, temperatures and humidities, especially in the lower levels, are considerably higher and convective instability extends to higher levels (note the winter and summer curves for Miami in figures 49 and 50). Furthermore there is a marked tendency toward increasing instability as the air moves over the warm land surface of the southern and central United States. This is in contrast to the tendency toward stabilization during the cold season, when the land is colder than the source region. The characteristic instability of Tg air in the summer persists even after it has traveled far to the north over the continent.

By the time it has reached the upper Mississippi Valley region though, it has lost so much of its moisture through showers, that the lift required to saturate the lower layers is generally so great that some type of frontal activity is necessary to cause thunderstorms. Thus, at times when Tg air occupies this region, ordinary thermal convection is insufficient to cause thunderstorms and only cumulus clouds or mild showers will be observed. With the passage of even a weak LOW, however, the forced ascent of the Tg air along the warm front, and later along the cold front, invariably causes severe thunderstorm activity. The fronts then provide sufficient lift to saturate the lower and intermediate levels of the air mass and to release the available convective instability.

Topography affects greatly the type of weather associated with Tg air. This is well illustrated in the arid, southwestern region of the United States. Thunderstorms may be expected whenever Tg air flows over this region in the summer. The lift produced as the air moves up over the mountains, combined with thermal heating in the afternoon, readily saturates the lower and intermediate levels and releases the convective instability which is characteristic of this air. These thunderstorms are a most important factor in the climate of this region, where a large portion of the rainfall occurs in the summer during the season of these showers.

Sonora Weather—These summer showers frequently are very intense and generally are extremely localized in their effects. The "cloudbursts" that occur in this region in the summer are almost

invariably caused by the presence of TG air either at the surface or aloft. Occasionally this type of air reaches the Pacific coast, and it is during these rare occasions that the southern California coastal region receives its unusual summer showers. The very severe cloudbursts that are occasionally recorded in the desert region of the southern Great Basin during the summer months are generally caused by an especially strong flow of TG air over this area. It is lifted by the high mountains and strongly heated during the afternoon so that, even though it has lost much of its moisture in traveling from its source region, nevertheless the combination of various lifting agents is sufficient to cause saturation and release of the available convective energy. Occasionally the TG air which reaches this area has traveled a very circuitous route across northern Mexico and approaches from the south instead of from the southeast. In this case the air may acquire additional moisture and instability in passing across the Gulf of Lower California. The synoptic situation for this *Sonora Weather* is the westward movement of the normal high-level anticyclone until it lies over the southern Great Basin. Under these circumstances TG air is fed directly from its source region to the southwestern United States and northwestern Mexico. The presence of a south*easterly* component in the winds above 6000 feet, instead of the normal southwesterly direction, will therefore cause the forecaster in this region to be on the lookout for this interesting type of summer weather.

In much the same way, TG air is lifted as it passes eastward across the Appalachian Mountains of the eastern United States, with the release of its convective instability and the production of severe local thunderstorms. With a strong southwest or west-southwest circulation involving TG air, therefore, severe thunderstorm activity may be expected in the mountainous regions of the eastern United States. These storms, which travel with the upper winds, frequently retain their activity after moving east of the mountains and appear on the coastal plains in the late afternoon or early evening, after traveling a hundred miles or more from the region in which they were formed. It should be noted that these thunderstorms are limited almost entirely to TG air. Polar air which returns northward after a 2 or 3 days' sojourn over the Gulf of Mexico rarely gives rise to this convective activity. It is thus very apparent that accurate identification of air mass types is invaluable in summer forecasting, just as it is during the winter. Since identification

presents more difficult problems during the summer, when contrasts at the surface are practically obliterated by the strong heating of that season, it is vital that the forecaster take particular care with this phase of the analysis. Tropical Continental air may frequently resemble TG air rather closely at the surface yet thunderstorms never occur in the former air mass while, as pointed out above, they are very common in the latter. Similarly, Superior air and returning Polar masses must be carefully distinguished from TG since thunderstorm activity never occurs in the first, and but rarely in the second type.

As TG air moves over the continent in the summer it frequently develops stratocumulus decks in the early morning. During the early or middle forenoon these generally break up into detached cumuli which may develop into thundershowers during the afternoon. Farther north, the early morning cloud decks are rare and they are replaced by occasional ground fogs. These are uncommon in fresh TG air but, not infrequently, occur in older masses in which convective activity has diminished and in which the lower levels have become polluted by smoke.

As TA air moves northward over the central and north Atlantic coastal region, it is cooled by contact with the cold water offshore and frequently develops sea fogs which move inland during the early morning hours. The formation of these fogs is considerably influenced by the wind velocity, for it has been observed that with comparatively rapid air movement they do not occur as frequently as with slow movement. This is probably due to the fact that with high winds the air moves across the belt of cool water lying off the east coast so rapidly that the lower portions are not cooled sufficiently to cause fog formation. With very light winds, on the other hand, the air movement may be so slight that any fog which forms offshore may not move inland more than a few miles and thus may affect only the immediate coast. (See page 309.)

Summer Flying Conditions in Tropical Gulf or Tropical Atlantic Air Masses—Stratocumulus clouds, with generally ample ceilings (500–2000 feet), are frequent in the Gulf coast region at night. During the day these clouds break into detached cumuli some of which usually develop into thundershowers during the afternoon. Farther from the source region the stratocumulus decks are rare and flying conditions are generally excellent after the cumulus activity decreases at night. After the TG air has been considerably

Average Tg Air in Summer Miami, Florida							
Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (3)	24	75	73	93	345	17.3	200
1000	20	68	....	88	346	14.9	300
2000	15	59	....	74	336	9.3	700
3000	9	48	....	58	333	6.3	900
4000	5	41	....	48	331	4.3	1100
5000	−1	30	....	54	333	3 4	1000

Average Tg_{1-2} Air in Summer Broken Arrow, Oklahoma							
Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (233)	30	86	69	58	348	15.4	1200
1000	27	81	....	50	345	12.3	1400
2000	20	68	....	55	342	9.9	1200
3000	13	55	....	62	339	8.2	900
4000	7	45	....	53	336	5.4	1000
5000	2	36	....	41	336	3.5	1300

Average Tg_{2-3} Air in Summer Ellendale, North Dakota							
Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg	Lift Meters
Surface (444)	29	84	71	66	351	16.5	900
1000	27	81	....	54	346	13 3	1200
2000	22	72	....	42	339	8.7	1600
3000	13	55	....	43	332	5.7	1500

Typical Tg_{2-3} Air in Summer San Diego, California 8–8–36							
Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (5)	22	72	71	95	338	15.3	200
1000	26	79	....	53	347	12.6	1300
2000	18	64	....	70	344	11.1	800
3000	9	48	....	93	342	9.9	200
4000	2	36	....	96	338	7.3	100
5000	−3	27	....	92	338	5.0	200

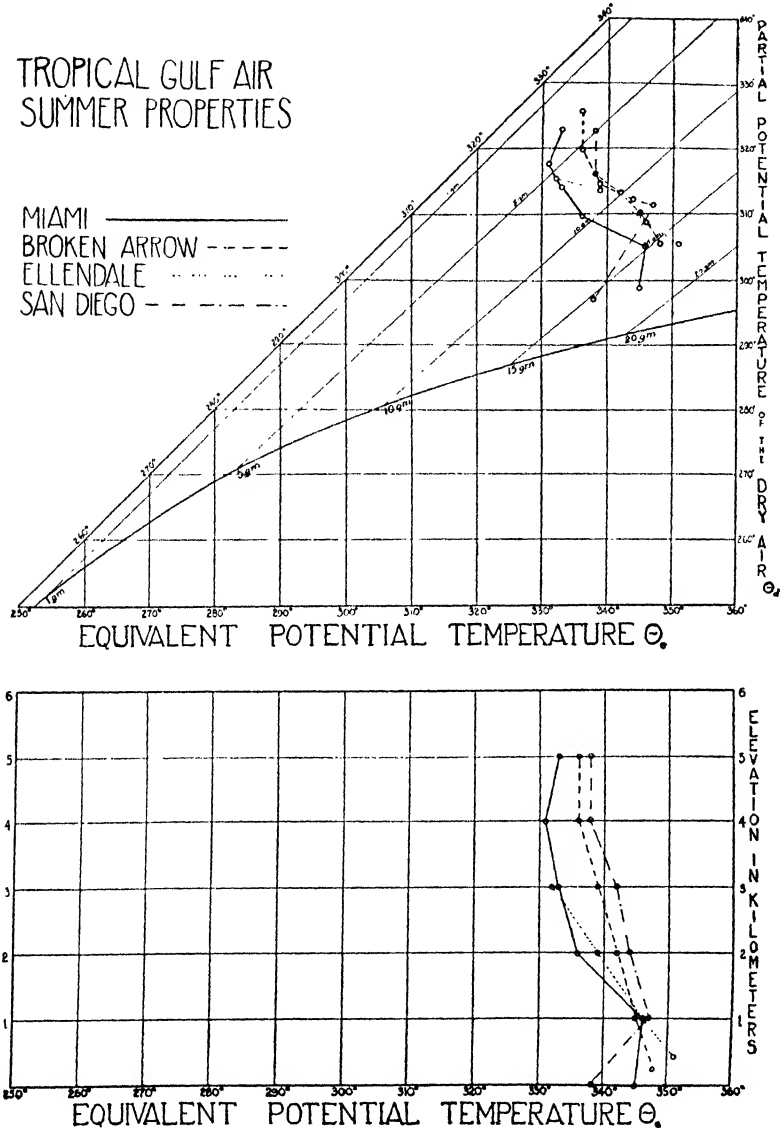

TROPICAL GULF AIR
SUMMER PROPERTIES
MIAMI
BROKEN ARROW
ELLENDALE
SAN DIEGO
PARTIAL POTENTIAL TEMPERATURE OF THE DRY AIR Θ_d
EQUIVALENT POTENTIAL TEMPERATURE Θ_e
ELEVATION IN KILOMETERS
EQUIVALENT POTENTIAL TEMPERATURE Θ_e

FIGURE 50

modified it may yield ground fogs. Cumulus activity within the air mass is always localized, with well scattered showers. Only in cases of frontal activity will showers coalesce to form large areas of bad weather.

Turbulence is considerable during the warmer part of the day when convection is active, especially near the surface. This roughness extends to high levels in actively developing cumuli, but can usually be avoided, due to the scattered nature of the showers.

Visibilities are good except when the air has moved far from its source and has become stabilized in the lower levels with development of fog. This is particularly noticeable along the north Atlantic coast, where sea fogs due to TA air are very common in the summer.

Characteristic Summer Properties of Tropical Gulf Air—During the summer, Miami also represents the TG source properties very closely. The remarkably close similarity between TG air and true equatorial air, as recorded at Batavia, Java, is apparent from a study of air mass curves from these two localities. Noteworthy in both cases is the marked convective instability, and the relative dryness aloft (cf. figs. 50 and 56).

As TG air moves northward over the continent, its instability apparently changes but little. Diurnal heating, however, leads to steep lapse rates in the lower levels and rapidly increasing instability during the afternoon.

Occasionally, as pointed out on page 152, TG air moves westward aloft to the Pacific coast region. An especially interesting case of this type of invasion is illustrated by the San Diego sounding for August 8th, 1936. Here TG air appears at all levels, even reaching down to the surface. This caused typical *Sonora* weather with oppressive heat and thunderstorms. The trajectory of this air took it over the Gulf of Lower California, where it acquired additional moisture in the lower levels. Perhaps some TP air also appeared in this outbreak.

TROPICAL PACIFIC AIR MASSES

The source region for this maritime air mass, which affects the Pacific coast of North America during the winter, lies in the high pressure belt of the North Pacific, between latitudes 25° and 35°. Surface water temperatures here are slightly lower (about 5° F.)

than in the same latitudes in the north Atlantic. The Pacific Anticyclone is, in general, better developed than the Azores-Bermuda HIGH, so it may be presumed that subsidence effects are more pronounced in the Pacific region. As far as the properties of the air from these two regions is concerned, after it reaches North America, the most important differences are due to the different trajectories involved. Tropical Atlantic and Tropical Gulf air which reaches the United States is derived from the *southern* limb of the north Atlantic HIGH. It thus comes directly from the warmest portion of the source region and is subjected to a relatively small amount of subsidence. Tropical Pacific air which reaches the west coast of North America, on the other hand, is derived from the *northern* limb of the source region and is subjected to a relatively large amount of subsidence. Furthermore, as Bergeron points out, air on the east sides of the subtropical high pressure cells is considerably more stable than on the west sides. (See page 88.)

It is interesting in this regard to compare the properties of Tp air on the Pacific coast of North America with Tropical Atlantic air as it reaches the coast of Europe. Such a comparison shows that the two air masses have almost identical properties, a fact which indicates that the source regions and trajectories must be very similar. Thus the marked differences between Tg and Tp air in the United States are due almost entirely to the fact that they are derived from different portions of their source regions, and not to any essential differences between the two sources. In fact Tropical Pacific air in the region west of Midway Island, and along the east coast of Asia is practically identical with Tropical Gulf air as it reaches the United States. (See page 177.) In both cases the air is drawn from the southwestern limb of a subtropical high pressure cell, from regions with almost the same surface properties.

Winter Properties of Tropical Pacific Air—During the cold season of the year Tropical Pacific air is of considerable importance in weather disturbances along the west coast of North America. It is brought into this region by low pressure areas which lie off the west coast of the United States and Canada, or over the Sierra Nevada and Great Basin provinces of the United States. This northeastward movement of Tp air from its source region between California and Hawaii is contrary to the general circulation around the Pacific HIGH. It only occurs when centers of action along the Polar Front move well to the south.

Tropical Pacific air differs considerably from Tg air as found in the central and eastern portions of the United States during the winter months. It is cooler, drier, and most important of all, much more stable. This is undoubtedly to be explained by subsidence effects in the Tp air masses as they pass around the Pacific HIGH. Because of this characteristic stability, precipitation resulting from Tp air is uniform, widespread and only moderately heavy, as contrasted to the heavy, squally type of precipitation which occurs in connection with Tg outbreaks. Squally weather in the Pacific coast and Great Basin regions is invariably due to a modified form of Polar Pacific air which possesses the necessary instability, and not to Tp air.

Tropical Pacific air is only rarely found at the surface over the continent and then only along the immediate coast during times when the general air flow is essentially parallel to the coast line. At other times, Tp appears aloft over the Pacific coast states, overlying modified Polar air. This feature of the occurrence of Tp air in the western United States is due to the rugged topography of that region. Any warm fronts involving Tp which approach from the west or southwest come to rest some distance offshore with the frontal surface lying on the mountain ranges which parallel the coast. The Tp air then pushes inland as an upper current, occasionally appearing at the ground at more elevated localities, but in general remaining well above the surface.

As the Tp air moves eastward across the Great Plains and Rocky Mountain regions, it remains aloft and can generally be identified with certainty only through aerographic soundings. The broad precipitation area often occurring in advance of occlusions which move across the Western States during the winter is usually caused by Tp air ascending the upper warm fronts of these systems (see figure 51). Only rarely can Tp air be identified east of the Rocky Mountain region. By the time it has reached the Great Plains it generally has been greatly dessicated by lifting over the mountains and over frontal surfaces, and resembles Superior air.

Outbreaks of Tp air occasionally may be followed well to the north into southern Canada, where they produce essentially the same weather conditions that they do to the south. After they have crossed the mountains into eastern Alberta and Saskatchewan, they are of comparatively little importance, since they have been dessicated at all levels by that time.

Tropical Pacific air may be distinguished from modified Polar Pacific air by its stability at all levels, and its comparative warmth aloft. Even greatly modified Polar Pacific air which has sojourned over the warmer waters of the south Pacific for from 6 to 8 days,

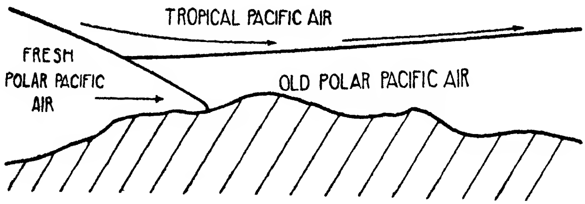


FIGURE 51.—PASSAGE OF COLD FRONT OCCLUSION OVER MOUNTAINOUS REGIONS OF WESTERN UNITED STATES

The Tropical Pacific air which overruns the upper warm front often causes widespread precipitation. The Polar Pacific air is marked by showery weather which generally clears rather rapidly.

retains its initial coldness at upper levels (above 3–4 km) and its characteristic instability. Superior air, which is probably only an upper phase of Tropical Pacific, is considerably warmer and drier at all levels due to intense subsidence effects.

Winter Flying Conditions in Tropical Pacific Air—Widespread precipitation usually accompanies the appearance of TP over the Pacific coast or Great Basin region. Flying conditions are often poor over a large area during such times, due to widespread rain and snow, together with low ceilings and poor visibilities. These weather conditions are all of a frontal nature, since TP air does not appear except in connection with an active warm front. Icing is generally only light or moderate in TP air due to its stability. It should not be confused with modified Polar Air in this regard for the latter may cause severe icing due to its instability and cold temperatures.

Tops of cloud systems in Tropical Pacific outbreaks are usually very high and the clouds are often arranged in layers with clear spaces between them. Over the higher mountains the clouds, however, generally form a solid deck from base to top. Turbulence is slight due to the stability of the air. Along the boundary layer, near frontal surfaces, it may be noticeable, however.

As TP air moves out aloft over the Great Plains it not infrequently passes aloft over a layer of cold Pc air and produces an extensive area of blizzard conditions. Under such conditions the clouds are relatively thin.

Average $_{2\text{-}3}$Tp Air in Winter San Diego, California							
Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (5)	19	66	55	68	316	9.1	700
1000	17	62	....	47	315	6.1	1400
2000	11	52	....	51	319	5.3	1200
3000	5	41	....	50	318	3.7	1200

Average Tp Air in Winter Honolulu, T. H.							
Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (3)	20	68	63	86	324	11.9	400
1000	15	59	....	81	325	9.8	400
2000	10	50	....	70	322	6.8	600
3000	6	43	....	51	320	4.0	1200
4000	1	34	....	33	320	2.1	1700
5000	−5	23	....	33	322	1.5	1700

Average Ep Air in Winter Coco Solo, C. Z.							
Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (3)	24	75	72	91	345	17.2	200
1000	19	66	....	89	341	13.9	300
2000	14	57	....	83	337	10.3	400
3000	9	48	....	54	329	5.5	1100
4000	4	39	....	50	329	4.0	1200
5000	−3	27	....	42	329	2.4	1300

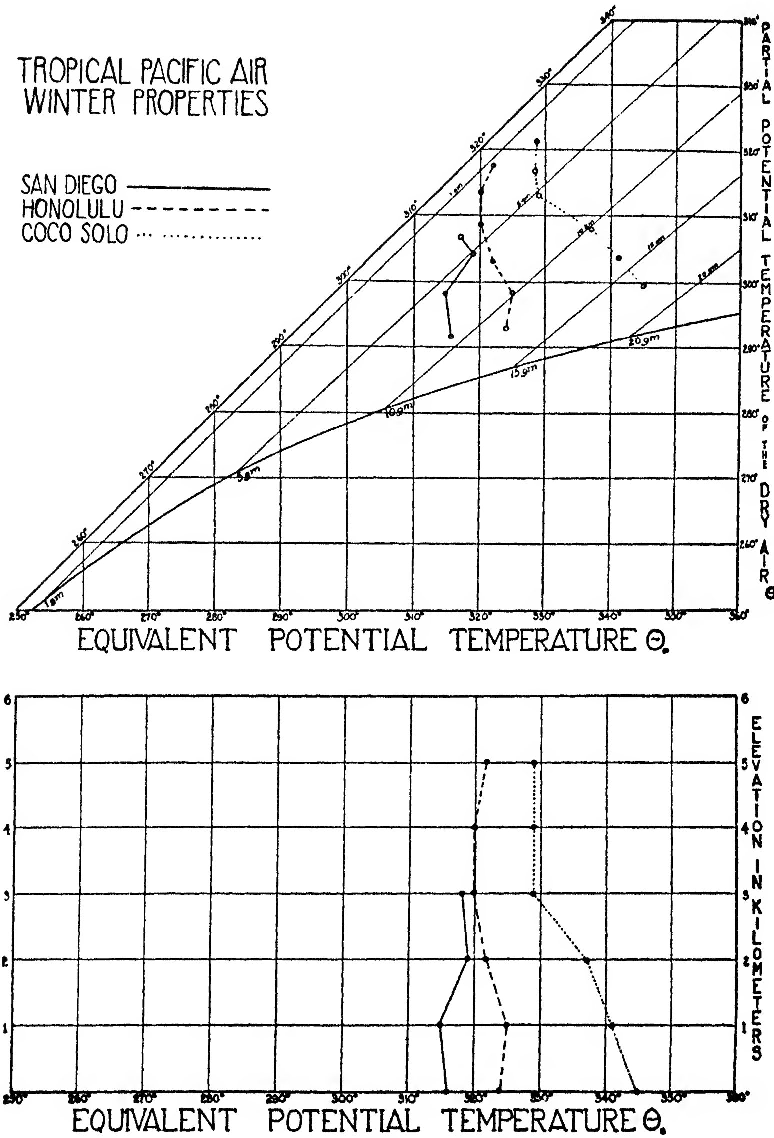

TROPICAL PACIFIC AIR
WINTER PROPERTIES
SAN DIEGO
HONOLULU
COCO SOLO
PARTIAL POTENTIAL TEMPERATURE OF THE DRY AIR Θ_d
EQUIVALENT POTENTIAL TEMPERATURE Θ_e
ELEVATION IN KILOMETERS
EQUIVALENT POTENTIAL TEMPERATURE Θ_e

FIGURE 52

Characteristic Winter Properties of Tropical Pacific Air—Soundings from Batavia and Honolulu give typical TP properties in its source region. Modifications as it reaches the west coast of the United States are shown by soundings from San Diego. These show considerable subsidence and a marked increase in stability from the TP source region properties.

Summer Properties of Tropical Pacific Air—Since the Pacific High is centered far to the north (35° to 40°) during the summer months and since the activity of the Aleutian LOW is very slight, Tropical Pacific air is unknown on the Pacific coast of North America during the warm season. It is likely that much of the air which reaches the coast during the summer and is classified as Polar Pacific air as actually initially of Tropical Pacific origin. This is so greatly modified by passage over the cool waters north of latitude 45°, however, that by the time it reaches the continent it is indistinguishable from PP air which initially comes from the Arctic regions.

Tropical Pacific air is of very great importance to the weather of the east coast of Asia since it plays much the same role there that T_G air does in the eastern United States. (See chapter 7.)

In the region of the north Pacific Ocean west of longitude 160°–170°, TP air is very important in the summer, causing fog, drizzle and rain as it moves northward over the cold water surface and is lifted by frontal activity. As it proceeds eastward around the Pacific HIGH, however, it is rapidly modified until it becomes indistinguishable from Polar Pacific air.

TROPICAL CONTINENTAL AIR MASSES IN THE UNITED STATES

From a study of the source region in North America available for the production of Tropical Continental air it is clear that these air masses will be of comparatively little significance. Actually, the only region available for the imparting of typical tropical continental properties, is the general area of north-central Mexico, western Texas and eastern New Mexico. Other tropical or subtropical portions of the North American continent, including central and southern Mexico and Central America, lie at high altitudes, are of small area and are of entirely negligible importance as source re-

gions for any air mass. Actually, air which may be classified as Tropical Continental consists of modified Polar Pacific or Polar Canadian air. This stagnates over the secondary source region of the southwestern United States and northern Mexico for several days becoming highly heated and dessicated in its lower levels. It then resembles in most important respects Tropical Continental air found in other portions of the earth. There is some tendency to include Tropical Continental air masses in the *Superior* air mass type described in the next section. However, when the origin of the warm, dry air which rather frequently appears over the south-central portion of the United States may be traced directly to the source region mentioned above, it seems desirable to classify it as Tropical Continental air.

This type of air, whether one considers it to be of true Tropical Continental nature, or merely to consist of somewhat modified Polar air, is of considerable importance in the weather conditions of the southwestern Great Plains and the Mississippi Valley during the summer. Conditions of severe drought in this region are almost invariably caused by the incursion of this air mass with its very high temperature, extremely low humidity and almost complete lack of any form of precipitation. It happens that these characteristics which are so disastrous to agriculture are especially favorable to flying.

Tropical Continental air is characterized by very low humidity, high day time temperature, large diurnal temperature range, considerable convective instability during the day and a very high condensation level. It appears only during the summer season since the source region is of no importance during the colder months of the year. It is readily distinguishable from Tropical Gulf air by its much lower humidity. Clouds and precipitation are almost entirely lacking due to the very high condensation level. However, strong surface heating often allows convective currents to reach rather high levels so that this air is frequently rather turbulent during the day. Occasionally, high level cumulus clouds appear in Tropical Continental air masses but the amount of moisture present is generally much too low to allow the formation of thunderstorms.

Surface air temperatures are generally very high during the day because of the great transparency of Tc air. It may frequently be distinguished in this manner from adjoining masses of Tropical Gulf air. The maximum temperatures in Tc air generally exceed

Typical Tc Air in Summer El Paso, Texas							
Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (1190)	24	75	56	52	341	11.0	1300
1500	27	81	....	37	344	9.7	2000
2000	24	75	....	43	347	9.9	1700
2500	23	73	....	36	344	7.8	2000
3000	18	64	....	43	344	7.6	1600

Average Tc_1 Air in Summer Broken Arrow, Oklahoma							
Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (233)	27	81	53	40	329	9.5	1700
1000	27	81	....	32	332	7.9	2100
2000	19	66	....	33	330	6.5	1800
3000	10	50	....	46	327	5.0	1300

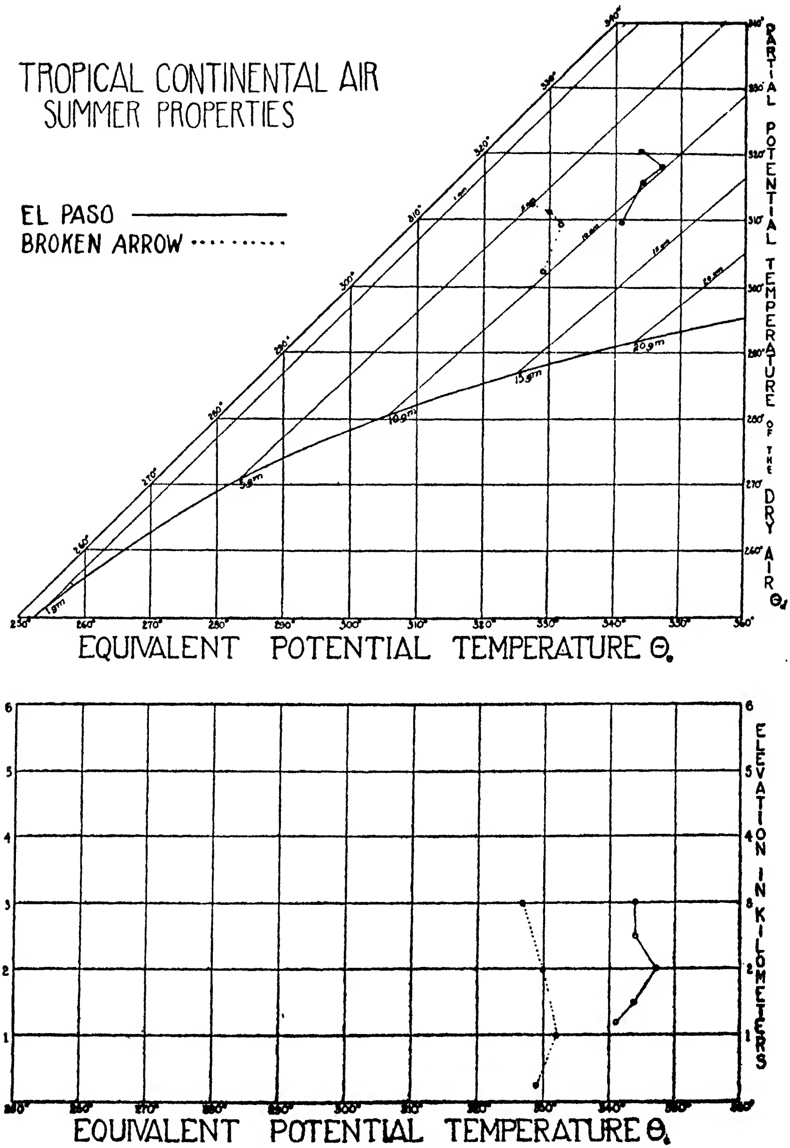

TROPICAL CONTINENTAL AIR
SUMMER PROPERTIES
EL PASO
BROKEN ARROW
PARTIAL POTENTIAL TEMPERATURE OF THE DRY AIR Θ_d
EQUIVALENT POTENTIAL TEMPERATURE Θ_e
ELEVATION IN KILOMETERS
EQUIVALENT POTENTIAL TEMPERATURE Θ_e

FIGURE 53

95° F., while the temperatures in adjoining Tropical Gulf air are usually from five to ten degrees lower. At night the Tropical Continental air masses cool off more rapidly than the Tropical Gulf air and may actually reach lower minimum temperatures.

The dew points in Tc air generally lie well below 50° F., and average between 30° and 40°. This is in very strong contrast to Tg air where the dew point in the summer time invariably lies above 60°. The boundary between Tc and Tg air masses may almost always be located by means of dew points. It is also clearly marked by the presence of thunderstorms in Tg air, and their absence in Tc air.

Extent of Area Affected—Since the Tc source region is very small the supply of this type of air is distinctly limited. Its outbreaks are frequent in western Texas, Oklahoma and Kansas, occasional in eastern Texas, Arkansas, Missouri and Nebraska, and very rare elsewhere. By the time it moves several hundred miles from its source region it is generally indistinguishable from other air masses that may be present, due to active mixing. It is practically never found at the surface east of the Mississippi River, and only rarely north of Nebraska.

Aerographic soundings in the southwestern Great Plains area frequently show a marked stratification, with distinct layers of warm and moist air. This occurrence of warm dry air at high levels above the earth's surface is generally explained by the presence of Superior air, since Willett has shown that these layers generally originate at high levels and are, therefore, quite distinct in their mode of origin from Tropical Continental air masses.

Flying Conditions in Tropical Continental Air—Clear skies or scattered small cumulus clouds are associated with Tc air. Visibilities are excellent due to the fairly steep lapse rate. Slight or moderate turbulence may be experienced near the ground in the afternoon, due to convective currents caused by strong surface heating, but this roughness may be avoided by flying well above the surface.

Characteristic Summer Properties of Tropical Continental Air in the United States—El Paso well represents the source region properties for Tc air. The air represented by the El Paso sounding (figure 53) was actually Pp air with a history of 3 days over water and 4 days over land ($_3Pp_4$) since leaving its source region. This air had become very much warmed in the lower levels as it

remained over the arid southwest. Finally it attained continental characteristics. Very high temperature, and high equivalent potential temperature are noteworthy. Also of interest are the low relative humidity, and rather complete mixing as evidenced by the lack of moisture stratification.

As this air moves northeastward over the Great Plains its properties usually change but little. As a general rule it tends to become slightly cooler and somewhat more moist. The properties of individual outbreaks vary within wide limits, however. The Tc properties for Broken Arrow thus differ from the El Paso properties in the somewhat lower temperature and humidity throughout. This difference is well within the rather wide variation in Tc properties.

SUPERIOR AIR MASSES

The warmest air observed over the United States is *Superior* air as defined by Willett. This air averages a few degrees warmer than TG air at intermediate levels and is somewhat colder than TG air at high levels, owing to the steep lapse rates frequently observed in it. Its relative humidity is almost invariably less than 40% and is frequently less than 20%. The specific humidity is also very low, ranging from 2 to 3 grams in the winter at elevations of 2 km, to 5 grams in the summer at the same elevation. At higher levels the specific humidity rapidly decreases, becoming less than 1 gram at elevations of 4 to 5 km.

Occasionally this air makes its way to the ground and then is only distinguished with difficulty from Tc air. Actually there is no essential difference in these air mass types at the surface except their trajectories. The Tc air originates at the surface over its source region in the southwestern United States and northern Mexico, while the Superior air originates in the upper levels of the atmosphere.

Willett believes that Superior air masses originate in the subtropics, as shown by their warmth, and that they have had a very extended past history in which they have remained aloft, as indicated by their extreme dryness. He believes that the source region is to be found in the upper levels of the subtropical belt of high pressure, and especially in the northern and eastern sides of the

great anticyclonic cells making up this belt. Thus, air from the Pacific HIGH slowly sinks as it moves eastward in the general circulation of the middle latitudes. In this way it is greatly heated by subsidence effects, with the result that its relative humidity is reduced to very low values. Due to its extended sojourn aloft its specific humidity is also remarkably low. These characteristics of warmth and dryness are intensified by foehn effects as the air flows down the eastern slope of the Rocky Mountains.

Superior air is generally brought into the United States region by WSW to WNW winds at intermediate levels. In the southern portion of the country Superior air may appear at nearly all levels in the lower troposphere. Farther to the north, where the polar currents are deeper, it appears at higher and higher levels and is rarely found at all north of latitude 50°. Since it appears chiefly in the upper levels of the atmosphere it is detected almost entirely by aerographic soundings. Its high temperatures and very low humidities are characteristic.

The weather conditions accompanying Superior air depend to a large extent on the depth of the current. When it occupies a large portion of the atmosphere below 15,000 feet comparatively little or no weather activity may be expected since the lift required to saturate it is so great that condensation forms are almost totally lacking. When Superior air is interstratified with TG air, with the Superior air predominant, it will cause a general diminution in convective activity. If, however, the Superior air only appears at high levels with a thick layer of TG air beneath, it may actually aid convection in the TG air due to its great dryness and freedom from cloud forms. Since rather steep lapse rates are frequently observed in Superior air, approaching the dry adiabatic at times, more or less turbulence is frequently noted in it at lower levels.

Generally speaking it appears only on cross-sections of the upper atmosphere since it rarely reaches the surface. When it does appear at the surface it often does so in the form of small tongues. These areas of dry air at the surface, in the midst of large areas of TG air, are often confusing to the analyst who considers only the surface indications. They are, however, readily explained if they are considered to be the surface manifestations of larger bodies of Superior air aloft.

This air mass type may occur at any season of the year. It is much more common, however, in the summer season, when Polar

air masses are relatively inactive. It rarely, if ever, appears at the surface except during the warmer season of the year. Outbreaks of Polar Basin air may occasionally be confused with Superior air masses since the properties of the two are somewhat similar. However, the synoptic situation leading to the production of Polar Basin air is very characteristic and should lead to an easy distinction between the two types. Furthermore, Polar Basin air masses do not exhibit the high temperatures characteristic of Superior air since they are of Polar origin, whereas the Superior air has remained in the tropics or subtropics for a number of days. Although the manner in which the source properties of both Superior and Polar Basin air are reached is quite similar—namely subsidence in a semipermanent anticyclone—the air involved differs greatly in the two instances and leads to the difference observed in the air mass properties.

Modifications of Superior Air Masses—The chief modifications of Superior air masses after they leave their source region result from mixing with other types of air. Superior air is of course comparatively little affected by surface conditions. As bodies of Superior air proceed to the higher latitudes they undoubtedly lose a certain amount of heat through radiational processes, so that their potential temperature may decrease somewhat. This is a rather minor effect, however, and in general they retain their initial properties for long periods of time. Most of the changes in temperature that occur are strictly adiabatic, and do not involve appreciable gain or loss in heat. The general modifications that are noted consist of a gradual increase in moisture as Superior air masses become mixed with TG masses. In localities which are considerably removed from the source region, practically all gradations from Superior to TG air masses may be noted.

Flying Conditions in Superior Air—Almost complete lack of cloud forms characterizes flying conditions in this type of air. Due to the tendency toward steep lapse rates the visibility is generally excellent. Slight turbulence may be experienced in regions of extreme surface heating but in general Superior air is smooth.

REFERENCES

1. Formation of Polar Anticyclones: H. Wexler. M. W. R. v. 65, n. 6, June 1937.

This is a most significant paper dealing with the processes which act during the production of a great mass of Polar Continental air. It is rather highly technical in treatment.

2. Cooling in the Lower Atmosphere and the Structure of Polar Continental Air: H. Wexler. M. W. R. v. 64, n. 4, April 1936.

This paper is along the same line of thought as the one cited above. It is a very important contribution to the whole, complex subject of the formation of great anticyclones.

3. The Air Masses of the North Pacific: H. R. Byers. Scripps Inst. of Oceanog., Tech. Series, v. 3, n. 14, 1934.

Byers presents a more or less preliminary study of the various types of air involved in cyclones over the north Pacific. He also includes several examples of the more common types of synoptic situations which occur in this region.

4. American Air Mass Properties: H. C. Willett. Mass. Inst. Tech., Papers in Phy. Oceanog. and Meteor., v. 2, n. 2, 1933.

This is the most important paper on North American air masses which has yet appeared. It discusses the various types of air, their source regions, their properties, and the types of weather which they produce. Characteristic curves, on the original Rossby diagram, are included for all of the more important air mass types.

5. Characteristic Weather Phenomena of California: H. R. Byers. Mass. Inst. Tech., Meteor. Papers, v. 1, n. 2, 1931.

This paper describes some of the more interesting weather types of the California region. It includes a chapter on the fogs of the interior valleys by W. M. Lockhart.

CHAPTER 7

AIR MASSES—ASIA

INTRODUCTION

Very little has been published concerning the air masses of Asia. The only detailed study of any kind has been carried on by H.-C. Huang. Mr. Huang's study, covering north China, is an admirable piece of work of great value to all meteorologists of the Far East. A. Wagner describes some upper air soundings from Batavia, Java, which yield valuable information on Equatorial air masses. The lack of weather maps, analyzed according to air mass principles, or aerographic soundings, from the Indian region precludes the obtaining of any information concerning the air masses of southern Asia.

AIR MASSES OF EASTERN ASIA

Polar Siberian Air Masses—The dominant air mass of the entire Asiatic region has its origin in the vast steppes of eastern Russia and Siberia. This is the largest strictly continental land area in the world. It is located in approximately the same latitude as Canada, so that air masses which originate in these two regions are very similar. They are of true Arctic nature, according to Bergeron's classification, grading into Polar characteristics toward the south.

H.-C. Huang has named air masses which have their origin in Russo-Siberia, *Polar Siberian* (Ps). They are characterized in the winter by intense coldness and dryness. No aerographic data are available from the source region itself but it can reasonably be assumed that there is no material difference between air originating here and in the Polar Canadian source region.

The only aerographic data available bearing directly on Ps properties are those which Huang obtained at Peiping. This is located near latitude 40° N., about 90 miles west of the Gulf of

Average Ps_0 Air in Winter Peiping, China							
Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (70)	−12	+10	−16	22	256	0.3	1700
500	−15	+ 5	....	24	261	0.3	1600
1000	−21	− 6	....	25	260	0.2	1500
1500	−25	−13	....	25	260	0.1	1400
2000	−30	−22	....	27	260	0.1	1300

Average Ps_1 Air in Winter Peiping, China							
Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (70)	2	36	10	29	277	1.2	1700
1000	− 6	21	....	30	277	0.8	1500
2000	−12	10	....	34	281	0.7	1300
3000	−17	+ 1	....	37	285	0.5	1100
4000	−23	− 9	....	41	289	0.4	900

Average Ps_2 Air in Winter Peiping, China							
Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (70)	14	57	28	31	294	2.9	1800
500	9	48	....	32	292	2.3	1700
1000	5	41	....	33	291	1.9	1600
1500	0	32	....	36	291	1.4	1400
2000	− 6	21	....	44	289	1.3	1100
2500	− 8	18	....	41	292	1.1	1100
3000	−12	10	....	58	293	1.3	600
3500	−15	5	....	63	296	1.2	500

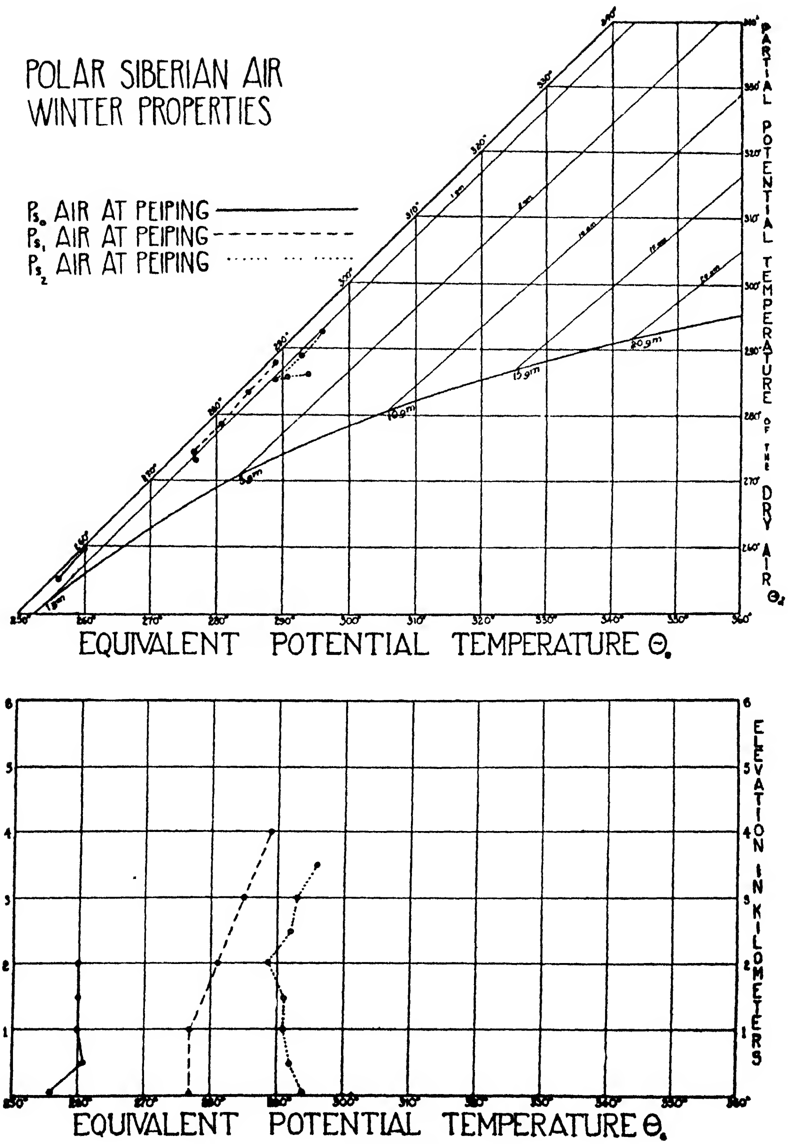

POLAR SIBERIAN AIR
WINTER PROPERTIES
P_{S_0} AIR AT PEIPING
P_{S_1} AIR AT PEIPING
P_{S_2} AIR AT PEIPING
PARTIAL POTENTIAL TEMPERATURE OF THE DRY AIR Θ_d
EQUIVALENT POTENTIAL TEMPERATURE Θ_e
5gm
10gm
15gm
20gm
ELEVATION IN KILOMETERS
EQUIVALENT POTENTIAL TEMPERATURE Θ_e

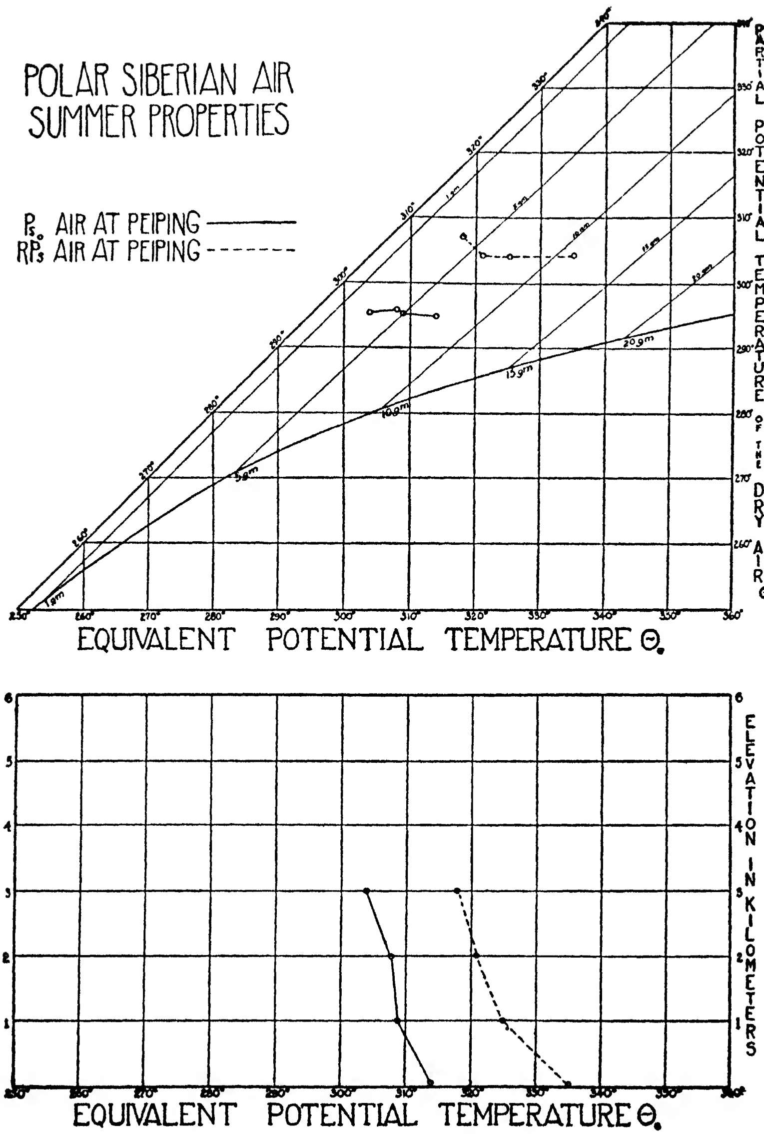
FIGURE 54

AVERAGE Ps_0 AIR IN SUMMER
PEIPING, CHINA

Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (70)	21	70	48	45	314	7.1	1300
1000	12	54	....	54	309	5.0	1000
2000	3	37	....	72	308	4.4	500
3000	−5	23	....	90	304	3.2	100

AVERAGE RPs AIR IN SUMMER
PEIPING, CHINA

Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (70)	30	86	60	41	335	11.1	1500
1000	20	68	....	47	325	7.6	1200
2000	11	52	....	58	321	6.0	800
3000	4	39	....	53	318	3.8	900



POLAR SIBERIAN AIR
SUMMER PROPERTIES
Ps₀ AIR AT PEIPING
RPs AIR AT PEIPING
PARTIAL POTENTIAL TEMPERATURE OF THE DRY AIR Θd
EQUIVALENT POTENTIAL TEMPERATURE Θe
ELEVATION IN KILOMETERS
EQUIVALENT POTENTIAL TEMPERATURE Θe

FIGURE 55

Average Tp Air in Summer Nanking, China							
Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (70)	29	84	76	78	360	19.6	400
1000	24	75	....	71	348	14 6	600
2000	16	61	....	71	337	9.9	500
3000	9	48	....	76	332	7.4	400
4000	2	36	....	73	331	5.3	400
5000	− 4	25	....	70	329	3.5	500

Average Ep Air—Yearly Average Batavia, Java							
Elev. Meters	Temp. ° C.	Temp. ° F.	D. P. ° F.	R. H. %	θ_e ° A.	W g/kg.	Lift Meters
Surface (10)	26	79	74	86	353	18.8	200
1000	21	70	....	77	342	13.3	400
2000	15	59	....	74	338	10.0	500
3000	10	50	....	66	335	7.1	600
4000	4	39	....	64	336	5.4	600
5000	− 2	28	....	60	338	3.8	700
6000	− 7	19	....	57	338	2.7	700
7000	−13	9	....	53	342	1.9	800
8000	−19	− 2	....	49	344	1.1	900

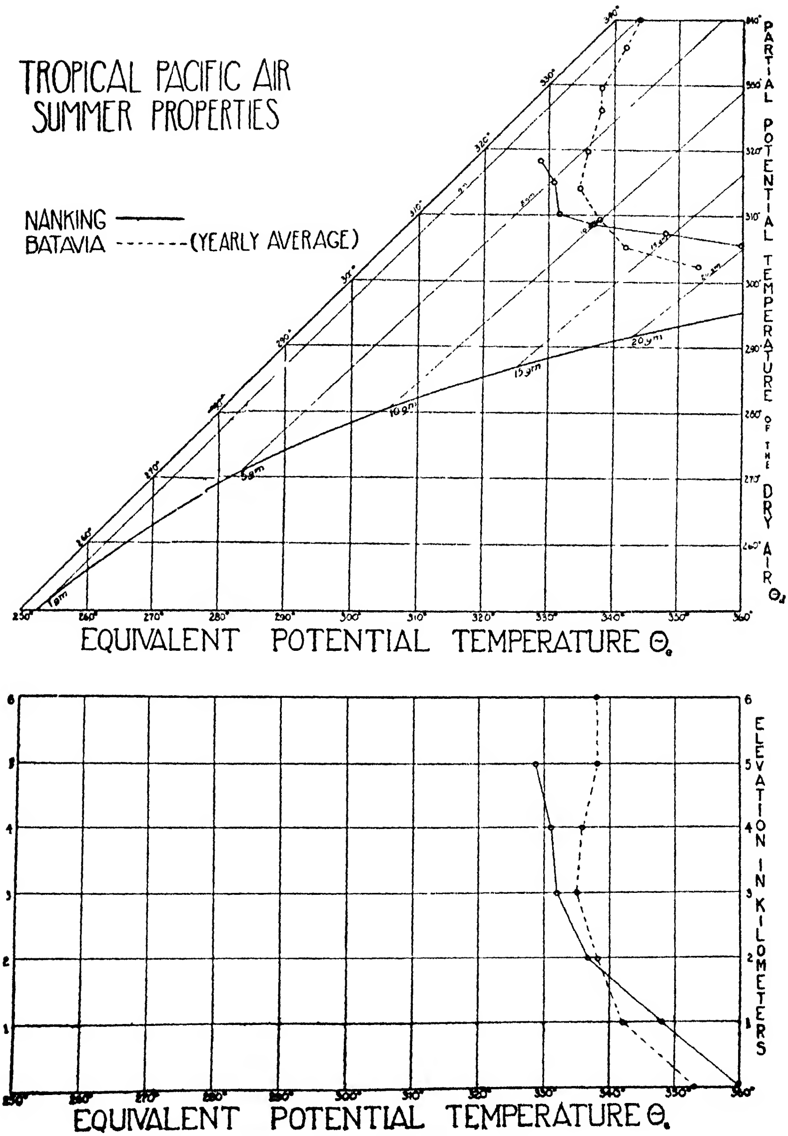

TROPICAL PACIFIC AIR
SUMMER PROPERTIES
NANKING
BATAVIA (YEARLY AVERAGE)
PARTIAL POTENTIAL TEMPERATURE OF THE DRY AIR Θ_d
EQUIVALENT POTENTIAL TEMPERATURE Θ_e
ELEVATION IN KILOMETERS
EQUIVALENT POTENTIAL TEMPERATURE Θ_e

FIGURE 56

Pohai, an arm of the Pacific Ocean. It is thus in approximately the same latitude as Indiana, so that Ps properties at Peiping should be very similar to Pc properties at Royal Center. A comparison of aerographic material from the two localities shows that resemblance actually is very close. Ps air is apparently somewhat drier than Pc air. It is also less stable. This is probably due to the higher average wind velocities at Peiping, which tend to cause mixing and destroy inversions.

Modifications in Ps air are shown in figure 54. The various continental trajectories, ranging from Ps_0 to Ps_2, were determined by Huang and represent average values for Peiping.

During the summer, Ps air at Peiping is also very similar to Pc air at Royal Center. At both localities the continental character is evidenced by the relatively low temperature and humidity. Actually, the similarity between the two stations is remarkable. Ps air at Peiping is somewhat more unstable than Pc air at Royal Center. This is probably due to the high wind velocities which prevail at Peiping.

Modifications of Ps air are described by Huang. He chooses air of various trajectories, ranging from fresh Ps air to greatly modified Ps air returning northward. The changes in air mass properties consist in the addition of heat and moisture and an increase in the instability.

Polar Pacific Air Masses—Only rarely does Polar Pacific air play a role of any importance in Asiatic weather. The general circulation from west to east in the higher latitudes precludes the possibility of Pp air reaching the coast of Asia except under rather unusual circumstances. With unusual development of the Siberian HIGH a return circulation of Ps air from the northeast, off the Bering Sea and the Sea of Okhotsk sometimes occurs. This produces weather phenomena along the north Asia coast analogous to those accompanying Polar Atlantic outbreaks over New England. A deep LOW, centered in the Japan Sea, may also cause the influx of Polar Pacific air over the continent. In general, however, Pp air is of small importance to Asiatic weather. Japan is influenced to a considerably greater extent by Polar Pacific air because of its maritime environment.

Tropical Pacific Air Masses—Air from the Tropical Pacific source region has a tremendous effect on the weather of China. As Huang points out, the periodic droughts and floods in the Yangtze

Valley are due almost entirely to variations in the supply of TP air in this region.

Aerographic soundings from Nanking, located near latitude 32° N., represent very well the slightly modified TP air which appears over the entire south China coast. This air is almost identical with TG air over the southeastern United States. In fact the resemblance between TP air at Nanking and TG air at Pensacola is most striking. In both cases the air is of true equatorial origin. It proceeds directly from the source regions with practically no subsidence effects. It is warm and moist at the surface and shows a rapid decrease in moisture aloft.

Tropical Pacific air near its source region is convectively unstable to a marked degree throughout the lower 4 to 5 kilometers. A lift of only 200–600 meters is required to saturate all of the air in this unstable region. For this reason the passage of TP air inland over the south China region in the summer is invariably accompanied by heavy showers.

Tropical Pacific air in the Asiatic region is very different from TP air which reaches the coast of North America. By the time TP air reaches North America it has been subjected to much subsidence in passing around the periphery of the Pacific HIGH. This tends to stratify the moisture and increase the temperature aloft, so that it becomes absolutely stable before entering into weather disturbances along the west coast of the United States.

Tropical Continental Air Masses—Apparently there are no Tropical Continental air masses of any importance to the weather of Asia. This is quite expectable, since the entire Asiatic continent lies well north of latitude 20° except for portions of India, Siam and Arabia. Furthermore, in the warm season, when Tropical Continental air might be formed, the summer monsoon forces tropical maritime air from the Indian Ocean and South China Sea over the entire southern portion of the continent.

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This is a rather general paper dealing with the types of air found over the China coast region. Little aerological material was available to the author, who was therefore able to write in only general terms.

2. Aerological Soundings by Kite-Flight at Peiping, China: H.-C. Huang. Bulletin of the Upper Air, Current Observations, Natl. Res. Inst. Meteor., Amademia Sinica, Nanking, 1935.

The author presents a fairly large amount of aerological material from the north China region, obtained throughout over 2 years. This is the first attempt at a comprehensive study of Asiatic air masses, and is very important. The author is writing a paper on "The Air Masses of North China" at the present time. This should furnish further valuable information concerning the air mass types of this region.

CHAPTER 8

AIR MASSES—EUROPE

INTRODUCTION

The air masses of Europe are in most respects similar to their North American counterparts. Polar maritime air as encountered in Germany (mAK) is thus very similar to Polar Pacific air at Seattle. Tropical maritime air as found in Germany (mTW), is almost identical to Tropical Pacific air at Seattle. It will be noted that local air mass designators fail in comparing air masses at widely separated localities. Thus the term *Polar Atlantic* means one thing in North America and a very different thing in Europe. For this reason it will be necessary to keep in mind the region *affected by the air* as well as its source.

Polar Atlantic air which affects North America is relatively unimportant. It actually consists of slightly modified Polar Canadian air that has had a short maritime trajectory. Air from the north Atlantic which affects Europe, however, is a most important air mass. If it is to be called Polar Atlantic here also, it must not be confused in any way with PA air in North America. It is, on the other hand, essentially similar to *Polar Pacific* air in North America. In this chapter, therefore, *local designators refer to air masses which act over Europe.*

The geography of Europe causes a profound difference in the relative importance of continental and maritime air masses in Europe and North America. Europe is open to direct invasions of very cold maritime air from the North Atlantic. North America, on the other hand, is well protected by high mountains bordering its west coast. The lack of large land areas to the north and northwest, on the other hand, allows cold continental air to appear relatively infrequently over Europe. Polar or Arctic Continental air is very common in North America, of course, since a large portion of the continent lies well above latitude 60° and serves as an important source region for cold continental air.

To the south, Europe is bordered by the great unbroken tropical land area of North Africa, a most important source region for tropical continental air masses. On the other hand, North America tapers rapidly southward so that tropical continental air is of very little importance here. Tropical maritime air reaches North America, especially the southeastern United States, directly from the source region. Any air of this character reaching Europe, on the other hand, is forced to travel around the Azores HIGH, with considerable alteration in its source properties.

Comparatively little has been written concerning the properties and classification of European air masses. Schinze's work in 1932 on this subject remains the only fairly comprehensive treatment. This work was in the nature of pioneering on the subject of air masses and already stands in need of revision. He described air masses largely by means of the thetagram, since charts similar to the Rossby diagram had not been developed then. As a result, it is difficult to determine the stability conditions of European masses from Schinze's data.

Schinze used a differential air mass notation similar to Bergeron's. This is inconvenient to employ in a discussion of local air masses. For this reason, a local classification, based on the source regions for European air masses, will be used in this chapter.

WINTER AIR MASSES

Arctic Atlantic Air Masses—The far northern Atlantic Ocean, between Greenland and Spitzenbergen, is the source region for the coldest maritime air found in Europe, if not on the globe. This air differs but little, in fact, from *continental* arctic air. It is extremely cold, very stable and contains very little moisture. It occurs frequently throughout the winter months and often covers much of the European continent.

This air mass generally appears over Europe during the passage of a deep low over the Scandinavian region. With this situation, air is drawn directly into the circulation from the source and reaches Europe very little modified from the source region properties. It is very cold, dry and stable. Although somewhat warmer and more moist than Pc air in North America, it is actually very similar to that air mass.

The exact trajectory taken by Arctic Atlantic air in reaching Europe is important. With a short path it exhibits its greatest coldness. With a longer path, its properties on reaching Europe approach closer and closer to Polar Atlantic air. The modification of the Arctic air over the ocean is rapid. With more than one or two days over the open ocean, the air acquires sufficient heat and moisture in the lower levels to become partially unstable. When this occurs it no longer may be termed true Arctic Atlantic air.

Polar Atlantic Air Masses—Air from the North Atlantic which has advanced into this source region from North America is termed Polar Atlantic air when it reaches Europe. It is essentially similar to the Polar Pacific air which reaches the west coast of

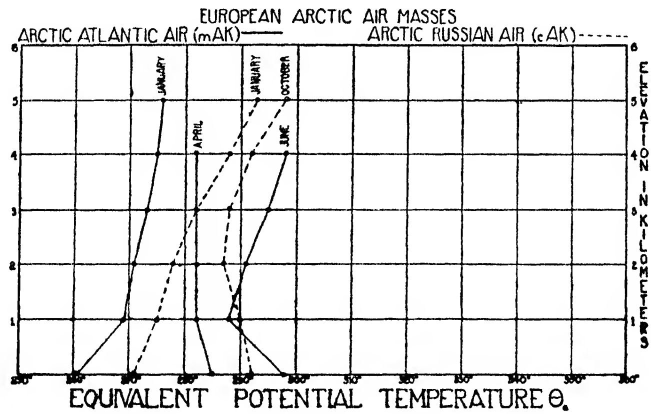

FIGURE 57

North America. Due to the open situation of Europe, however, this air proceeds practically unchanged until it reaches the Alps. Thus the entire continent of Europe from the Atlantic Ocean eastward to the Alps enjoys an essentially maritime climate much of the time. Polar Pacific air masses which reach the interior of North America, on the other hand, have lost so much moisture in ascending the Rocky Mountains, that they show but slight indication of their maritime origin.

Polar Atlantic air as it reaches Europe is either in approximate neutral stability or is somewhat convectively unstable. It thus yields mild showers of the air mass type. When lifted by frontal activity, however, the air often gives rise to heavy showers. When it strikes the Alps it is rapidly lifted a considerable distance. This

is often the first important lifting to which it is subjected after leaving its source. In such cases it causes unusually heavy precipitation in the mountains.

The air which is modified to produce Polar Atlantic air may be either Pc air which has crossed part of the United States or Canada or Polar Pacific air which has crossed the entire North American continent. In some cases, old Tropical Gulf air may be modified to the Polar Atlantic type in crossing the northern Atlantic Ocean. Most of the outbreaks of Polar Canadian or Polar Pacific air which reach the east coast of North America are involved in well occluded cyclones. These are almost always regenerated in the general zone of frontogenesis and cyclogenesis along the western north Atlantic

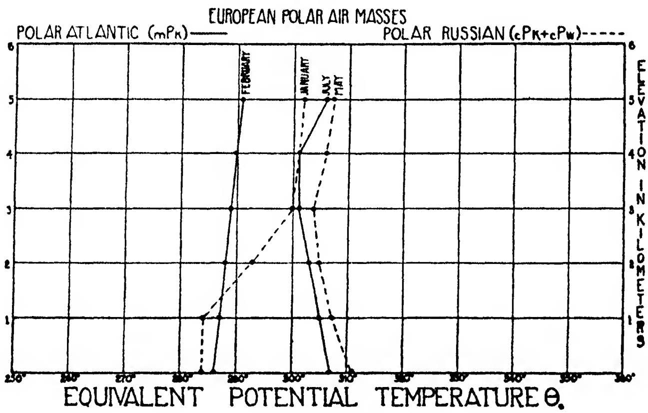

FIGURE 58

Ocean. They then proceed eastward across the Atlantic Ocean and reach Europe in connection with only slightly occluded waves.

Arctic Russian Air Masses—This type of air is rather infrequent over Europe. The only source region for it is northernmost Russia. It reaches Europe under the rather infrequent pressure distribution in which an intense HIGH lies well north of Russia and a deep low lies over Scandinavia. This air mass type is very important to central and eastern Asia (Polar Siberian), but plays a very minor part in European weather.

Arctic continental air is essentially similar to Pc air in North America. It is stable, cold and dry. As it reaches central Europe it is not particularly cold, however, since it has moved a considerable distance from its source. For this reason Europe does not suf-

fer with the extremely severe cold waves that frequently overspread all of the central part of North America during the winter.

Polar Russian Air Masses—Western Russia and Finno-Scandia furnish the source region for this important European air mass. It occurs fairly frequently during the winter over much of Europe. It is rare during the summer, for at that time the deep LOWS necessary to draw it over central and western Europe do not occur. It is stable during the winter months and resembles, fairly closely, modified Pc air as encountered over the south-central United States during the cold season. It is considerably warmer and drier than Arctic air masses which reach Europe.

Tropical Atlantic Air Masses—These air masses are essentially similar to Tropical Pacific air which reaches the Pacific coast of the United States. They are warm, moist, and show a slight increase in equivalent potential temperature aloft. They are thus stable both for the dry and the saturated states. When lifted along frontal surfaces they yield widespread, rather even rains. The showery conditions found in Tropical Atlantic air as it reaches the southeastern United States do not occur over Europe. Subsidence as the air travels around the periphery of the Azores HIGH, causes this notable change in structure.

Tropical Atlantic air often affects the entire European continent. Widespread winter rain and snow storms with the precipitation sector covering most of western and central Europe are usually the result of activity of this air mass.

Tropical Saharan Air Masses—This interesting continental air mass type is of considerable importance to the weather of Europe and southwestern Asia. The great tropical desert region of northern Africa is one of three similar areas over the world which are capable of producing important masses of warm, dry air. Of the other two areas, northern South America is much more humid and modifies maritime air masses comparatively little. Central Australia is very similar to the Sahara region but air from here affects only the Australian continent itself. The major portions of North America and Eurasia lie well above the tropics and no part of either one is of importance as a continental tropical source region.

In the winter, Tropical Saharan air is relatively warm, dry and markedly stable, with a rapid increase aloft of the equivalent potential temperature. It enters the European region under the influence of well developed LOWS which move across the southern part

of the continent. It brings good weather with clear skies and high temperatures. Marked contrasts frequently occur when this air is displaced by Polar Atlantic air behind cold fronts, which cross the

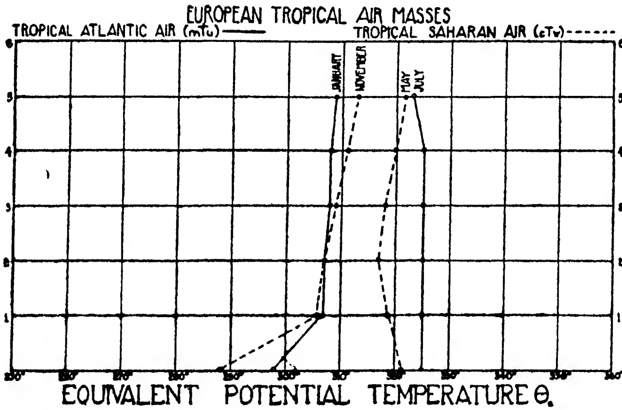

FIGURE 59

European regions. It very frequently moves aloft over central and northern Europe after crossing the Alps, appearing as a warm dry current from the southwest.

SUMMER AIR MASSES

Polar Atlantic Air Masses—During the warm season, the northern air masses of Europe are all essentially of Polar character rather than Arctic. There is little difference then in the source properties of the north Atlantic between the British Isles and Canada, and to the far north near Greenland.

Considerable variation occurs in the properties of air masses originating in this large area. Air which reaches Europe directly from the northern portion with a relatively short maritime trajectory exhibits mild instabilty with a slight decrease aloft of θ_e in the lower levels. Air which arrives from the southern portion of the source region after a fairly extended sojourn over the ocean is generally stable throughout and shows a slight increase aloft of θ_e. The unstable form of Polar Atlantic air commonly gives rise to mild convective activity. When forced aloft by frontal activity it may cause rather severe showers. The stable form gives only scattered cumulus

clouds under most situations. Its condensation level is rather high so that it requires a considerable amount of lift to produce precipitation. Only when it is involved in well developed LOWS does it act as a rain producer.

Tropical Atlantic Air Masses—Only rarely does air from the Tropical Atlantic source region reach Europe in the summer without considerable modification. During the warm season the Azores HIGH is usually well developed. As a result most of the air from the south Atlantic must pass far to the north, over the cold waters of the north Atlantic, before reaching Europe. When it does reach the European regions the air is stable and not very warm. The equivalent potential temperature is practically constant at all levels. It causes active precipitation only when involved in well developed cyclones. Nothing comparable to the thunderstorm conditions which accompany the influx of Tropical Gulf air into the southern United States is observed in Europe.

Tropical Saharan Air Masses—The summer weather of the entire continent of Europe is greatly influenced by this warm and dry air mass. The passage of even very weak low pressure centers across central or northern Europe causes it to flow northward across the Mediterranean into Europe. It is slightly convectively unstable in the lower levels, and then stable above 3 kilometers. Its condensation level is so high, however, that rarely does it cause more than a few cumulus clouds.

The conditions that bring about the outflow of air from the Sahara region usually also cause severe dust storms there. These result in the transport of dust over the European region. As the Saharan air moves farther and farther away from its source the larger dust particles settle out to the ground. The smaller particles often remain in suspension for very long periods giving to this air a characteristic opalescent haze. This is even recognizable in Tropical Atlantic air which has had its origin in the Sahara region.

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2. Troposphärische Luftmassen und vertikaler Temperaturgradient: G. Schinze. Beitr. z. Physik d. freien Atmos., Bd. 19, 1932.

These two papers constitute Schinze's pioneering work on a classification and description of European air masses. They are remarkably complete for their time, but they are not up to date now. Comparatively little is said about the instability conditions of the various air mass types, and very little indeed about the synoptic situations under which they occur. These papers are both entirely in German.

CHAPTER 9

FRONTS

INTRODUCTION

Whenever two air masses with different properties approach one another a zone of transition is formed in which the properties of one merge into those of the other. Since it is along these zones of separation between adjacent air masses, or *fronts*, that much of the weather activity in the temperate zones is observed, their study is of great importance to the meteorologist in these regions. The extent, both vertical and horizontal, and the width or "sharpness" of fronts vary within very wide limits. Generally speaking it may be said that a front must extend above the levels affected by purely surface phenomena, perhaps 3000–5000 feet, and have a length of at least 200 miles in order to be of synoptic importance. The width may vary from about 3 miles to perhaps 50 miles. Fronts cannot be much less than 3 miles wide because of turbulence, which causes mixing of the air on the two sides even in the sharpest examples. When they exceed about 50 miles they are strictly *frontal zones* and represent a gradual change of the properties of one air mass to those of the other. Such frontal zones may represent either the conditions which precede the formation of an active front or those which occur after an active front has diminished greatly in activity and the air masses on the two sides have begun to mix. It is important to keep in mind the order of magnitude of fronts since it will keep the meteorologist from making the error of considering minor or localized zones of transition, such as are found along coastal regions, as true fronts.

In many instances it will be found that the properties of one air mass merge into those of another along a zone, perhaps 300–1000 miles in width. In such cases it is frequently difficult to locate properly the actual position of the front. If it is remembered that the front always represents a distinct *sharpening* of this zone it will generally be possible to locate it accurately on the ground. In case

no such sharpening of the zone is present it is clear evidence that no true front exists.

DISCONTINUITY SURFACES

Figure 60-a represents the conditions along a front of infinitely small width, a theoretically perfect *discontinuity surface.* A property of the two air masses involved, such as temperature, is represented by the isotherms t_0, t_1, . . . t_5. These isotherms are distinctly offset along the discontinuity surface. t_2 in air mass A is thus located at some distance from t_2 in air mass B, t_3 is some distance removed in the two air masses, etc. There is thus a very sharp change in the temperature along the discontinuity surface. This abrupt discontinuity in air mass properties is never attained in the atmosphere, as pointed out above, because of turbulence effects.

Figure 60-b represents a distribution of temperature which might occur naturally along a very sharp front. Here the transition from the temperatures observed in A to those in B takes place across the front FF′, here shown to be very sharp, perhaps 5 miles wide. The temperature distribution in A is typical of a fresh polar air mass with a steep gradient. That in B is typical of a warm tropical air mass with a much less steep gradient. The gradient across the front is extremely steep. In actual cases it occasionally amounts to 20°–30° F. in 5 miles. Figure 60-c represents the temperature conditions along a more diffuse front, perhaps 50 miles wide. Here the temperature gradient across the front is much less than in 60-a or 60-b and the contrast between the two air masses is less. In this case the temperature difference across the front may amount to but a few degrees, and the two air masses may differ but little in general properties. Such a situation is common during the summer during weak outbreaks of Polar air which may differ but little from the air they displace.

Finally, figure 60-d represents the conditions along a general transition zone in which no sharpening in the temperature field appears as air mass A yields to B. No front is in existence but rather a general zone of mixing. In all of the cases mentioned, temperature was used to illustrate the change in properties along a front. Actually, humidity, wind direction and other properties change in a similar manner.

Although, as pointed out, fronts have a definite width, ranging from a few miles up to 50 miles or so, for most practical purposes

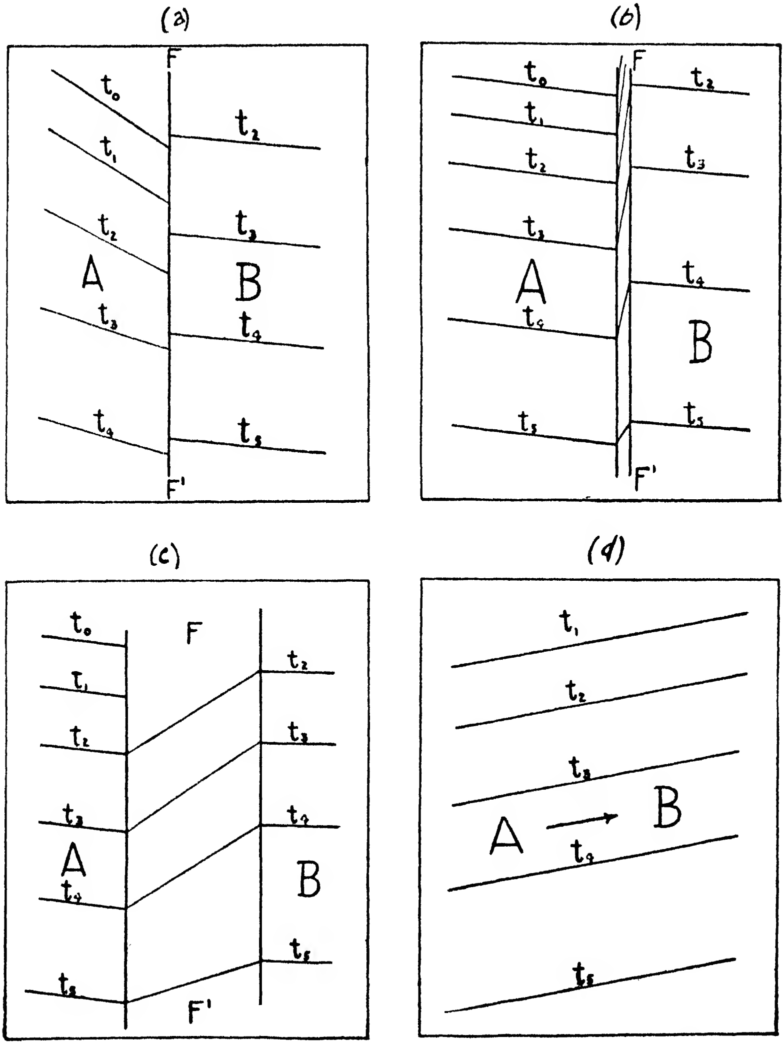

FIGURE 60.—TYPES OF DISCONTINUITY SURFACES

they may be considered as definite discontinuity surfaces in which the properties of the air mass on one side pass abruptly to those on the other. They may then be treated according to the ordinary

laws of hydrodynamics. It is found that the slope of fronts may be expressed by the following equation:

$$\tan b = \frac{d}{g} \frac{(t_2 v_1 - t_1 v_2)}{(t_2 - t_1)}$$

where b is the slope of the front, d the Coriolis force (the deflective force due to the rotation of the earth), g the acceleration of gravity, t_1 and t_2 the temperatures, and v_1 and v_2 the velocities of the air currents on the two sides of the frontal surface. This assumes that the warm and cold currents are flowing past one another side by side in opposite directions. For equilibrium conditions it may be shown from this equation that: (1) the cold air must underlie the warm air, (2) the smaller the temperature contrast the steeper the front, (3) the greater the difference in velocity of the two currents the steeper the front. It is assumed, in deriving the above equation, that the air in the two currents flows in a straight line—this is not strictly true, but this assumption does not change any of the general results.

In summary, for equilibrium conditions between two air currents of different temperatures: (a) the colder current must underlie the warmer and flow toward the right of an observer standing in the cold air mass, and (b) the smaller the temperature difference or the larger the velocity difference, the steeper the slope of the boundary surface.

TYPES OF FRONTS

Rarely does it occur that a discontinuity surface between two different air masses will remain stationary for any great length of time. Generally one of the air masses tends to move forward and displace the other, with the result that the discontinuity surface or front will also move. Stüve has pointed out four major types of moving fronts. These are classified according to the following outline, where the upper air mass—

1. flows *actively upward* along the front,
2. moves *passively upward* along the front due to active underrunning and lifting by the cold air beneath,
3. flows *actively downward* along the front,
4. moves *passively downward* along the front due to active recession of the cold air beneath.

The first two of these frontal types are the well known warm and cold fronts of Bjerknes. The latter two types are generally referred to as *surfaces of subsidence,* rather than fronts, since they represent zones of very little meteorological activity. They commonly occur in anticyclones.

Warm Fronts—The type of front in which warm air flows actively upward over a colder wedge beneath, generally has a rather gentle slope of from 1–100 to 1–400. It is generally characterized by relatively small contrasts in temperature and wind direction when compared with the cold front, chiefly because the actively advancing warm air has not been heated above normal prevailing temperatures, to the extent that cold air is frequently cooled.

Because of the very gentle slope of the warm front it may extend its influence far in advance of its surface position, giving rise to cloud systems as much as 1000–1500 miles away. The warm front cloud system is frequently very complex and may involve practically every common cloud type (figure 61-a). First precursors of the distant front are generally cirrus clouds. These are probably not connected directly with the front, but are caused rather by a wave along a high level inversion in the upper troposphere which is induced by the distant cyclonic circulation. These cirri are not simply detached tufts nor are they of the *cirrus densus* type. They are, rather, the well defined *cirrus radiatus* or *cirrus filosus* types, with a very definite banded arrangement. Several hours after the first cirri have been observed, the first clouds directly connected with the warm front will be seen. These are of the *cirrostratus filosus* and *c. lenticularis* forms which exhibit marked arrangement in filaments and bands. They generally do not have the same orientation as the preceding cirri.

It is still open to considerable question whether the cirrostratus cloud deck mentioned above is the result of overrunning along the warm front itself or whether it is caused by some form of activity along an upper front surface above the warm front. When viewed from an airplane flying at an elevation of 12,000 to 15,000 feet, in the cirrostratus zone of the warm front activity, it is seen that the high cirrostratus clouds do not merge with the intermediate clouds which follow them but rather continue as a solid deck out over the intermediate clouds toward the warm front. Actually there are probably several upper fronts which mark the levels of each of the principal cloud layers. These probably merge with the main warm

front as shown in figure 61-a, with the main front marking the boundary below which no activity occurs.

Considerably below the cirrus and cirrostratus clouds are found altostratus and altocumulus types at elevations varying from about 8000–15,000 feet. These occur in well defined decks which do not merge into the cirriform clouds but rather lie at considerably lower altitudes. These cloud types are the first ones which are definitely related to the main warm front surface. The clouds in this region vary greatly, with their form depending on the type of air which

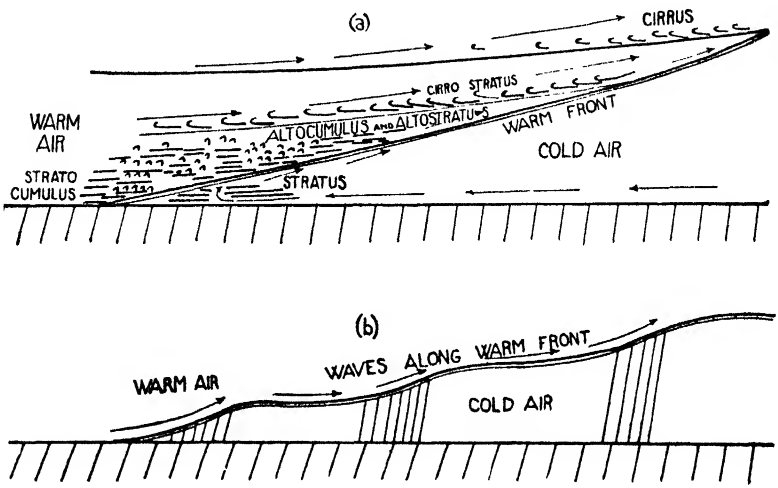

FIGURE 61.—(*a*) CLOUD SYSTEM ACCOMPANYING TYPICAL WARM FRONT. (*b*) WAVES ALONG WARM FRONT SURFACE, PRODUCING IRREGULAR PRECIPITATION

is overrunning the warm front. If this air is comparatively stable a flat layer of altostratus will be formed; if it is somewhat unstable a turbulent deck of altocumulus will result; if it is extremely unstable a heavy bank of *altocumulus castellatus* with towering cumuli above will be formed, with perhaps the production of a line of active thunderstorms. In the latter case scud clouds may form in the precipitation areas, whereas in the first two cases, lower clouds are generally lacking in this region.

As the warm front surface approaches closer to the ground, the entire cloud system increases in thickness and the various intermediate and lower levels thicken and merge, whenever the frontal

activity is sufficient to cause precipitation. Unless the front produces precipitation the altostratus or altocumulus clouds described above are generally the only intermediate or low forms that are present. Low clouds, when they occur in connection with warm front activity, may be the result of two major causes. They may appear just above the front surface as the result of lifting and condensation there, or they may occur in the cold air mass beneath the front, due to lifting of the cold air as a result of convergent airflow near the intersection of the frontal surface with the ground. The production of low stratus decks by this method is described in detail on pages 311-315.

Although observational data are too scanty to afford definite conclusions, it is probable that the surfaces of warm fronts may frequently have a distinct wave form with the crests of the waves parallel to the intersection of the front with the ground. This is indicated by weather phenomena scarcely explicable in any other manner, particularly those involving irregular precipitation. It is occasionally noted, for example, that areas of rainfall under a warm front show a variable intensity that is apparently related to varying rates of overrunning. At least in some instances the situation shown in figure 61-b must be of marked influence in controlling the rate of precipitation falling from a warm front.

The passage of a warm front at the surface is marked by a number of weather phenomena, including:

1. an abrupt temperature rise,
2. a slight wind shift, usually about 45°,
3. a slight barometric trough,
4. a slight isallobaric discontinuity,
5. a rapid clearing in weather conditions,
6. a rapid increase in specific humidity.

Flying Weather—An airplane or airship flying through a warm front aloft will generally experience the following weather conditions:

1. an abrupt change in temperature: falling if the path is away from the surface position of the front, and rising if toward it,
2. a wind shift amounting to 30°–90°,
3. slight to moderate turbulence,
4. in some instances, high level thunderstorms,

5. a marked change in cloud types and elevations,
6. a marked change in icing conditions (see chapter on Icing).

Cold Fronts—The second type of front as defined by Stüve is one in which cold air actively displaces warm air, causing it to be forced upward. The weather phenomena accompanying this front are very different from those observed with warm front. The primary difference, perhaps, is that in the case of the cold front

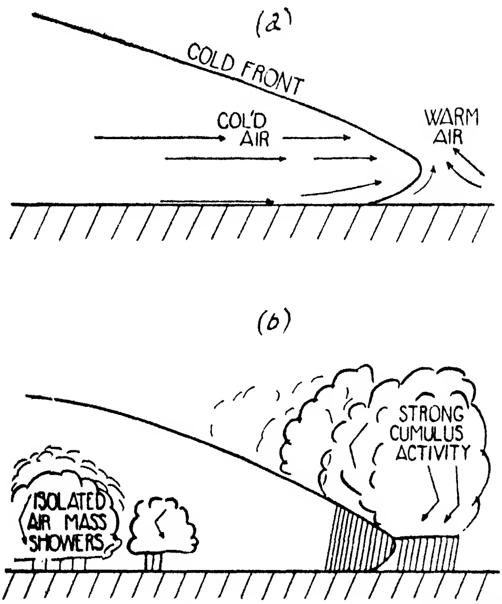

FIGURE 62.—GENERALIZED CROSS-SECTION OF COLD FRONTS
(a) Airflow. (b) Weather phenomena.

the meteorological activity is confined to the immediate vicinity of the front, instead of being distributed over a wide zone in advance of it. This is largely because of the much steeper slope of the cold front, which ranges from 1–30 to 1–100. Since the front slopes back away from its direction of motion, practically no cloud forms or other weather phenomena herald its arrival. Furthermore, since it generally moves considerably faster than the warm front and since its slope is much steeper, the area of ascending air marking the region of bad weather near the front is narrow and passes very rapidly.

The forward edge of the cold front in contact with the ground tends to be retarded by friction with the result that the front frequently takes on the shape shown in figure 62. The portion of the cold air which pushes out in advance of the main air mass is obviously in a very unstable position since it is denser than the air beneath it and tends strongly to sink to the ground and displace the warmer air. This is precisely what occurs along an actively advancing cold front and accounts for the strong turbulence that is frequently observed in such a case. The well known *Line Squall* or *Windshift Line* with the extremely turbulent conditions that it exhibits, is nothing more than a well developed cold front.

As long as the front moves steadily and rapidly, clearing behind it takes place promptly. When the front, however, is subjected to retardation for any reason, clearing may not occur for a considerable time. In fact, a low stratus or stratocumulus cloud system may form due to convergence in the air under the front. This type of weather phenomenon is discussed in some detail on pages 311-315.

The passage of a cold front at the surface is marked by many weather phenomena, including:

1. an abrupt temperature fall,
2. a well marked wind shift, amounting to from 45°–180°,
3. a barometric trough, generally very well marked,
4. a very pronounced isallobaric discontinuity,
5. a general improvement in weather conditions, sometimes very rapid, sometimes gradual when the air behind the front is unstable. Occasionally no improvement may result due to convergence immediately following the front. Whether the improvement is to be rapid, slow, or will not occur at all, may generally be decided by a study of the air mass properties of the cold air, and from a consideration of the pressure and tendency fields.
6. a marked decrease in specific, and also generally, of relative humidity.

Flying Weather—An airplane or airship flying through a cold front will generally experience the following weather conditions:

1. an abrupt change in temperature: falling if the path is away from the surface position of the front, and rising if toward it.

2. a wind shift amounting to from 45°–180°. The wind-shift will be most pronounced near the surface and will decrease in magnitude with elevation.
3. moderate to strong turbulence,
4. in many cases, thunderstorms,
5. a marked change in cloud types,
6. if temperature conditions are suitable, moderate or severe icing.

Secondary Cold Fronts—If the temperature gradient is considerable within the cold air mass involved in a cyclone, conditions will be favorable for the formation of secondary cold fronts. These generally occur during times of outbreaks of unusually cold Polar air which is moving over a warm land surface. The secondary fronts may occur at intervals of several hundred miles and each may be marked by a minor pressure trough, tendency discontinuity, temperature differential, slight wind shift, and general cold front weather phenomena (see figures 81 and 85). The early recognition of these secondary fronts is important for aeronautical purposes since the general squally weather conditions which occur within a fresh mass of polar air are usually concentrated along these secondary fronts. They are commonly separated by regions of comparatively good weather. The development of secondary fronts rarely occurs during the warm season of the year for at this time the strong temperature gradients within the air mass necessary for their formation are not present.

Secondary cold fronts often form when a cold body of air passes from a continental region over an adjacent warm ocean surface. This is noted during the winter, as polar Canadian outbreaks pass eastward over the Atlantic Ocean from North America. During this season secondary cold fronts form within the cold air, or those already present, are at least considerably intensified. A similar situation often occurs along the borders of the Great Lakes during outbreaks of Polar Canadian air. The intensely cold air is somewhat warmed in its lower levels as it passes over the warm water surface. This causes the formation of secondary cold fronts along the northern borders of the lakes. These fronts sweep southeastward with the general circulation and cause heavy snow flurries and very poor flying weather as they pass. During a strong Pc outbreak a continuous series of these secondary cold fronts, sepa-

rated by intervals of 6–12 hours, may pass across the region between the Great Lakes and the Appalachian Mountains for several days after the initial outbreak. This type of activity diminishes during the latter part of the winter as the more shallow lakes become completely frozen.

Secondary warm fronts are practically unknown since the temperature gradients in warm air masses are almost always very slight. There is practically never sufficient concentration of isotherms to cause frontogenesis.

Occlusions—When a cold front approaches and meets a warm front, the warm air lying between them must be forced upward and outward. This displacement of the air in the warm sector is an important step in the development of an extratropical cyclone and is known as the *occlusion process.* It may take place in two different ways as shown in figures 63 and 64. In the first case, the air behind the advancing cold front is the densest of the three air masses involved in the cyclone; while in the second case, the cold air behind the cold front is denser than the air immediately in front of it, but lighter than the cold air beneath the warm front. Thus, the wedge of cold air in the first place will displace both of the air masses in front of it (figure 63). In the second place it will displace the warm air immediately in front of it, but will pass up over the cold air beneath the warm front (figure 64).

The former process is called a *cold front type occlusion,* while the latter one is called a *warm front type occlusion.* Both of these types are of great importance in daily forecasting, and an early recognition of which type is occurring is essential, since the type of weather accompanying them differs greatly.

Cold Front Type Occlusion—The weather which accompanies a cold front occlusion is of a composite type. Prefrontal clouds and precipitation result from activity along the upper warm front (figure 63). Thus the cold front type occlusion is preceded by all of the phenomena which precede the common warm front. Cirrus and altostratus cloud decks usually herald its approach. Widespread prefrontal precipitation is common, although the area affected by it is generally smaller than with a well developed warm front. Scud clouds and stratocumulus decks may form under the altostratus layer as the front approaches, as with a true warm front. It is possible for convergent activity in the cold air beneath the warm front to occur preceding the arrival of the front, but this is relatively un-

common, since the necessary convergent airflow is usually lacking.

As may be seen from figure 63, a cold front occlusion acts at the surface in exactly the same manner as a cold front. Therefore the weather accompanying and following the two types of fronts is practically identical. A well marked wind shift line, pressure

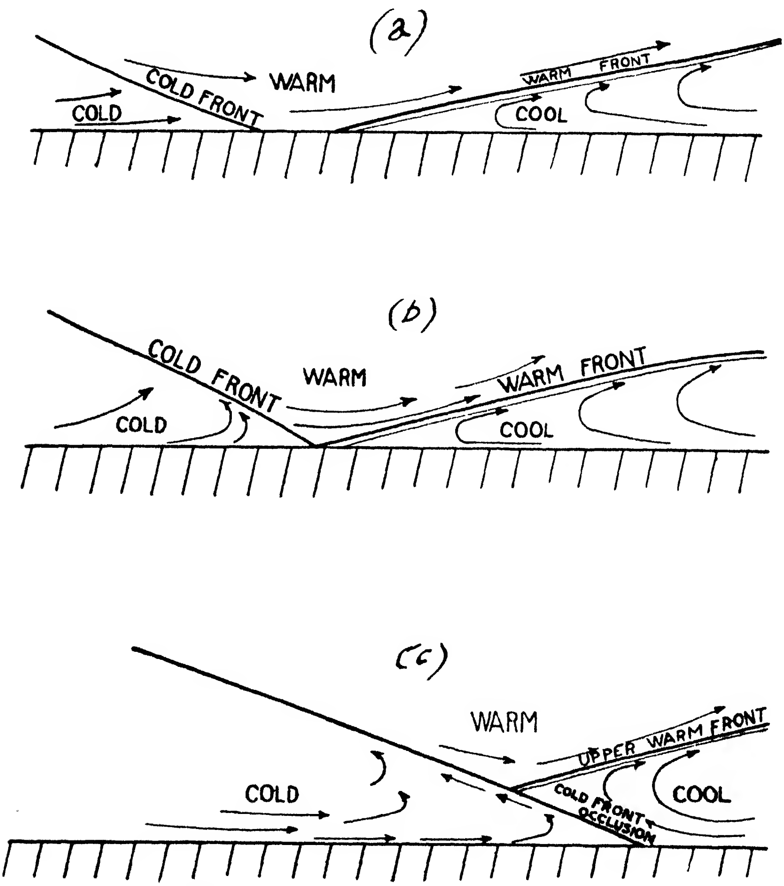

FIGURE 63.—COLD FRONT TYPE OCCLUSION

trough, tendency discontinuity, temperature and humidity contrast are all present in both cases. Instability showers are common in the air behind the cold front occlusion when it is convectively unstable. Convergent activity following the passage of the cold front occurs relatively frequently.

Flying Weather—The zone of poor weather preceding a cold front occlusion is usually not over 200–300 miles wide and it is rare that definitely unflyable weather precedes the front by more than 100 miles. The most unfavorable weather will be found in the immediate vicinity of the front, where true cold front phenomena may be expected. Following the passage of the front a

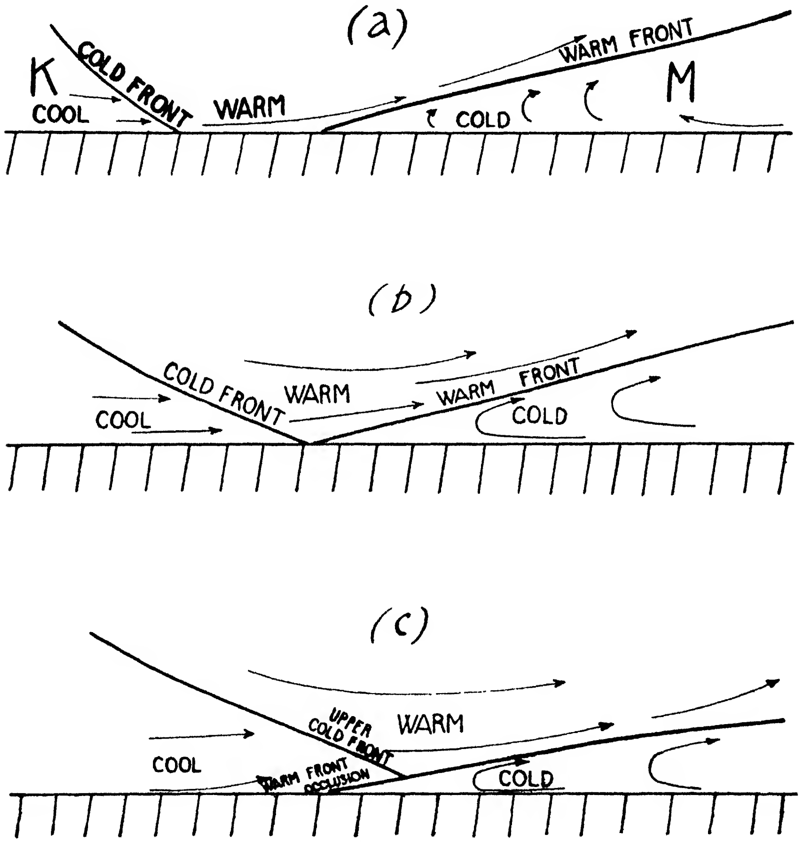

FIGURE 64.—WARM FRONT TYPE OCCLUSION

marked improvement in flying weather usually occurs, unless the front is being retarded. In such cases convergent cloud systems may hamper flying considerably within a zone 50–200 miles wide.

Warm Front Type Occlusion—Whereas the cold front occlusion described above exhibits both cold front and warm front characteristics, the warm front type occlusion acts in most regards

like a warm front. Because of the great difference in the weather which accompanies these two types of occlusions it is very important to distinguish them in their early stages. In many cases this is very difficult since the difference in density of the two cool air masses (figure 64) may be only slight.

In general, the weather accompanying the warm front type occlusion is that of a typical warm front with a few minor differences. Of especial interest is the rather heavy, showery type of precipitation that frequently occurs when the upper cold front (figure 64) passes over a region. This showery weather is not ordinarily encountered in the case of ordinary warm front activity although it may occur when the overrunning air is rather unstable. In the case of showers in connection with a warm front, however, they are irregularly distributed over a wide region in advance of the surface position of the front. Showers which are caused by the upper cold front of a warm front type occlusion occur in a rather narrow band. This band moves across the region under the warm front surface, more or less parallel to the surface position of the occlusion. (See figure 65-b.)

It will be noted that the band of showery weather is limited to definite areas, between x and z (figure 65-a). In the area between the surface position of the front and x, the overrunning warm air has not received sufficient lift to saturate it. In the area beyond z, the air has been lifted so far that most of its moisture has been removed and the passage of the upper front is marked only by a band of clouds. Both of these factors must be given careful consideration when forecasting. If aerographic soundings are available the width of the rainless areas between the front and x may be determined by calculating the slope of the warm front and computing the lift necessary to saturate the overrunning air.

Along mountainous coastal regions where the surface position of the warm front lies at some distance out to sea, and where the overrunning air is rather moist, the rain area will commonly extend to the immediate coast line and even some distance seaward. In cases where the warm front occlusion occurs over relatively flat country, and where the overrunning air is somewhat drier, the rainless area may be about 100 miles wide. The position of the area beyond z varies principally with the slope of the warm front surface. The steeper the slope the closer it will be to the surface front. It is also dependent on the moisture distribution and stabil-

ity of the overrunning air. The more stable and the drier the air aloft is, the closer it will be to the surface front.

The warm front type occlusion is particularly common along the windward coasts of mountainous continents or large islands. In such locations cold fronts which sweep in from sea frequently bring air which is colder and denser than the air it is displacing at sea, but which is warmer than the air lying along the immediate coastal

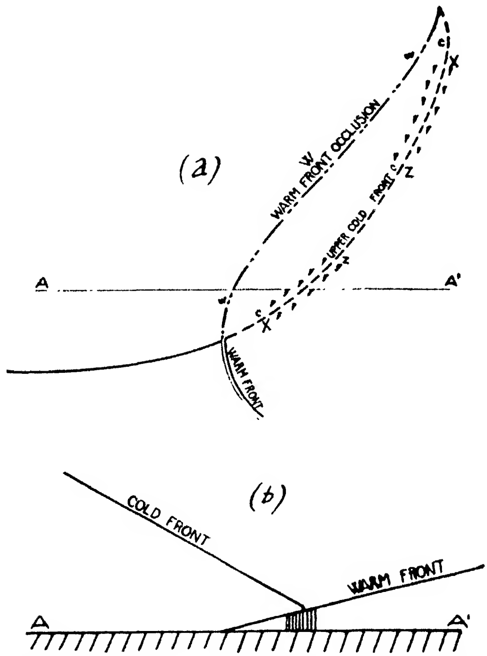

FIGURE 65.—DETAILS OF WARM FRONT TYPE OCCLUSION

(a) Plan view, showing showery weather along upper cold front. (b) Section, showing precipitation area.

region. Furthermore, the cold air along the coast may be trapped between mountains which are parallel to the coastline and the surface of the warm front, as shown in figure 66. In such cases, if the cyclonic activity is sufficient to cause a continental supply of warm, moist, maritime air, precipitation may continue for a considerable period of time. The heavy rainfall which occurs along the west coast of North America, north of latitude 40°, is largely to be attributed to this situation.

The warm front type occlusion is also very common in the area east of the Rocky Mountains during the winter. At such times the entire Great Plains region may be occupied by very cold Polar Canadian air. Due to cyclonic activity in the general region of the northern Great Basin or southwestern Canada, Polar Pacific air may be drawn across the Rocky Mountains behind a cold front. This air may be colder and denser than the air it displaces in the Great Basin-Rocky Mountain region, but it often is considerably warmer and lighter than the cold Polar Canadian air which lies

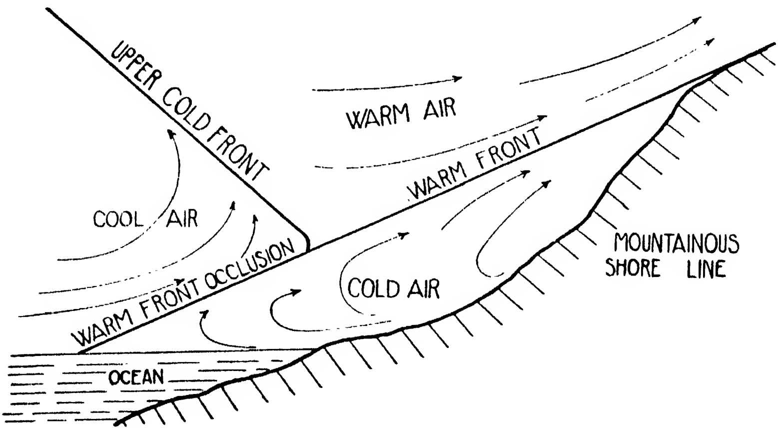

FIGURE 66.—STRUCTURE OF WARM FRONT TYPE OCCLUSION ALONG MOUNTAINOUS COAST

The cold air which is trapped beneath the warm front acts as a wedge to force the warm air aloft.

just east of the mountains. A warm front occlusion thus may occur a short distance east of the base of the mountains. The Polar Pacific cold front then proceeds out aloft over the Great Plains region. This situation is described in some detail in the section on Polar Pacific air masses.

The early recognition of the warm front type occlusion is important and frequently rather difficult. Several criteria may be employed to distinguish it from the cold front type occlusion and from the true warm front and it is important that these be applied to all doubtful cases.

1. If the approaching cold wedge of air, marked *K* in figure 64-a, is of nearly the same surface temperature or actu-

ally warmer than the air beneath the warm front (marked *M*), the occurrence of a warm front type occlusion should be suspected.

2. Particularly along coastal regions, the wind shift along the warm front occlusion (*w-w-w* in figure 67-a) is often not of the magnitude to be expected with a cold front type occlusion. This criterion does not hold, however, in inland localities where the wind shift is generally well marked in both cases.

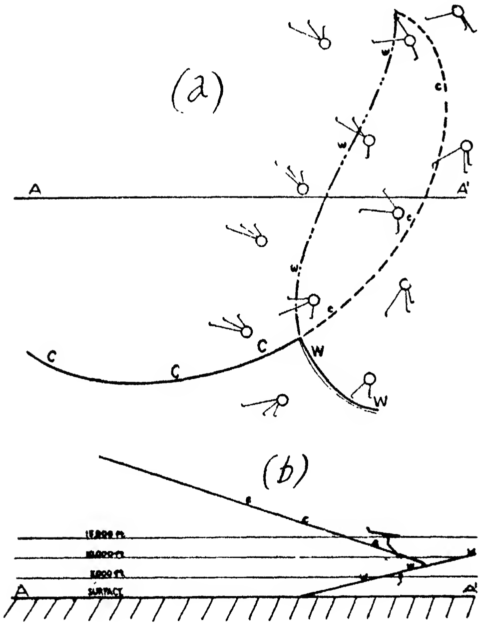

FIGURE 67.—DETAILS OF WIND FLOW IN A WARM FRONT TYPE OCCLUSION

(*a*) Plan view. Short arrows, winds below 5000 feet; medium arrows, winds from 5000–9000 feet; long arrows, winds from 9000–13,000 feet. (*b*) Section along line AA′.

3. The presence of an isallobaric discontinuity some distance in advance of the surface position of the occlusion is perhaps the most reliable means of locating the position of the upper cold front and thereby identifying the occlusion as one of the warm front type. Whenever an occlusion occurs about which there is any question the synoptician should examine very carefully the barometric tendencies in advance of it with this in mind.

The tendency distribution shown in figure 68 is very characteristic of a warm front type occlusion and should be studied carefully so that it may be recognized when encountered in making a map analysis. Here it will be noted that a distinct tendency discontinuity exists along the warm front type occlusion. However, on further inspection of the tendency field it will be noted that along the line (*c-c-c*) another weaker tendency discontinuity exists. This marks the position of the upper cold front and is in many cases a most important feature of the analysis.

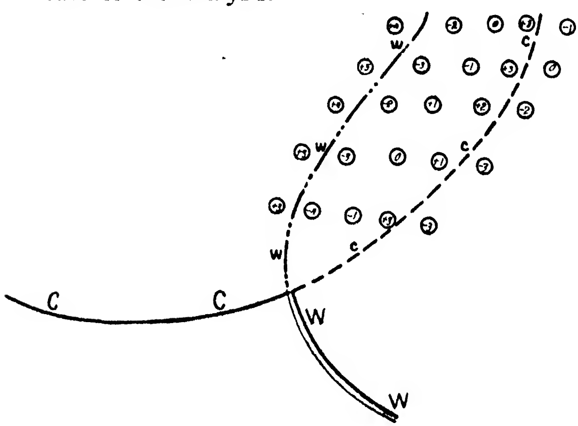

FIGURE 68.—TYPICAL PRESSURE TENDENCY DISTRIBUTION IN A WARM FRONT TYPE OCCLUSION

Note the tendency gradients across both the warm front occlusion at the surface and the upper cold front.

4. When upper wind information is available it may be employed very usefully in determining upper fronts. This criterion must be used with caution, however, and only in connection with other data, since other phenomena than the warm front type occlusion may produce a similar velocity field. In figure 67 the upper wind field is shown with arrows. The short arrows represent average winds below say 5000 feet, the medium arrows winds from 5000 to 9000 feet, and the long arrows, winds from 9000 to 13,000 feet. In common map practice these arrows are drawn in different colors, so that the features which they indicate may stand

out clearly. A windshift in the lower levels (strictly the surface winds) marks the position of the warm front occlusion at the surface (*w-w-w*). A slight windshift also marks the surface position of the warm front (*W-W*). The most interesting portion of the velocity field lies in the region between the surface position of the warm front occlusion (*w-w-w*), and the position of the cold front aloft (*c-c-c*). Here a marked discontinuity in the lower level winds (those below 5000 feet) and the intermediate level winds (those between 5000–10,000 feet) exists. This upper discontinuity represents the position of the warm front occlusion aloft as shown in figure 67-b. Occasionally a discontinuity may be observed between the intermediate and the high level winds, although this is rarely seen, since the clouds in the vicinity of such a system generally preclude the obtaining of sufficient upper wind data.

Flying Weather—The flying conditions change unusually rapidly with a warm front type occlusion. In the earliest stages the occlusion acts as an ordinary warm front with the usual cirrus or cirrustratus and altostratus cloud decks. Even these may be lacking if the overrunning air is especially dry. The first unique weather is the sudden appearance of a band of showers as the air behind the upper cold front reaches saturation. Ceilings in this shower area are generally fairly high (above 2000 feet). However, areas of low scud clouds may be present if the shower activity is pronounced. Most important for the pilot is the frequent occurrence of icing along the upper cold front.

Once the band of showers along the upper cold front is established, it moves steadily along with the upper front. After traveling several hundred miles it usually loses intensity as the overrunning air reaches such a high elevation that most of its moisture has been extracted.

Generally fine weather may be expected near the surface position of the occlusion within a short time after the cold front moves aloft. Prior to the occlusion process this region may be affected by warm front precipitation. Activity ceases in nearly every case, however, as the cold front passes and displaces the warm air. Often this improvement is rather sudden, after a long period of unfavor-

able weather. If the air behind the cold front is rather moist, however, no improvement at all may occur. This is frequently the case along coastal regions where precipitation may continue as one upper cold front after another passes up over a well established warm front occlusion. This occurs only in the situation shown in figure 66 where the cold wedge of air is held stationary by the terrain. Normally, the cold wedge slowly recedes after the occlusion process has started.

Composite Upper Fronts—A rather unusual, though interesting, type of warm front occlusion occurs when a *cold front occlusion* is involved in the situation shown in figure 69. The cold front

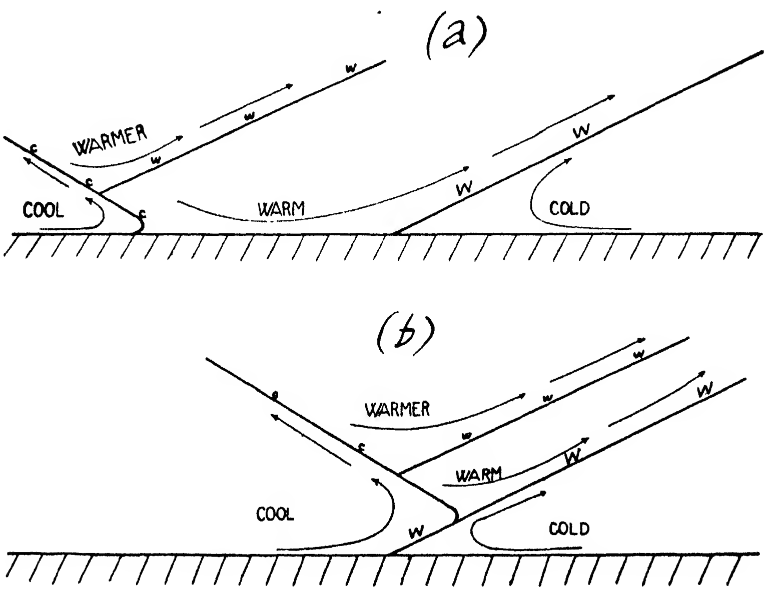

FIGURE 69.—COMPOSITE WARM FRONT TYPE OCCLUSION WITH AN UPPER COLD FRONT OCCLUSION

occlusion (*c-c-c*) is shown moving toward the warm front (*w-w-w*) in figure 69-a. In 69-b the cold front has passed upward along the warm front in a warm front type occlusion. There are now two warm fronts involved in the occlusion. Usually the air above the upper warm front (*w-w-w*) is so dry that little activity occurs there, although the additional lift provided by the occlusion process is often sufficient to cause a cloud deck (altostratus) and occasional high level precipitation.

This phenomenon occurs occasionally east of the Rocky Mountains when old occlusions move up over cold polar air at the surface. Pilots have described upper cloud systems and precipitation areas at such times that closely fit the theoretical situation.

REFERENCES

1. Synoptic Determination and Forecasting Significance of Cold Fronts Aloft: B. Holzman. M. W. R., v. 64, n. 12, Dec. 1936.

This is an excellent paper on the recognition and importance of upper cold fronts. It is well illustrated.

2. Upper-Air Cold Fronts in North America: Stephen Lichtblau. M. W. R., v. 64, n. 12, Dec. 1936.

This paper presents further examples of upper cold fronts.

Also see papers noted under chapters 10 and 11.

CHAPTER 10

FRONTOGENESIS

INTRODUCTION

The importance of air mass boundaries, or fronts, in weather phenomena has been described in the previous chapter. It is the purpose of the present chapter to point out some of the means by which regions of front formation may be recognized on the weather chart. Recent research by Petterssen has done much to clarify this important subject. His work will form the basis of this chapter.

Particularly in forecasts for 36–48 hours, it is very important to recognize as early as possible the appearance of fronts and frontal zones. Petterssen has shown that this is easily possible from a study of the wind and temperature distribution over a region. The basic elements of his theory and its practical application will be described below.

FRONTOGENESIS

Frontogenesis is a process which concentrates the solenoid field between two air masses. This concentration is the result of the pushing together of equiscalar surfaces—such as those representing potential temperature—into a narrow zone. During frontogenesis, the curves of constant potential temperature move so that they tend to produce a discontinuity along a line. This line is called the *line of frontogenesis.* It may be straight or curved, stationary or moving.

If the property (potential temperature) which is being subjected to frontogenesis be denoted by α, then its ascendant (or vector representing its greatest rate of increase) will be given by $\nabla\alpha$. The magnitude of this vector will be represented by $|\nabla\alpha|$. If frontogenesis is to take place, $|\nabla\alpha|$ must be of *greater* magnitude along the line of frontogenesis than elsewhere. In case $|\nabla\alpha|$ is of *smaller* magnitude along a certain line than elsewhere, *frontolysis* is occurring and the line represents a *line of frontolysis.* Since

frontolysis is the exact opposite of frontogenesis it will not be considered separately.

Changes in the configuration of the equiscalar surfaces α to produce frontogenesis are produced by atmospheric movements. These may be represented by the *velocity field,* which is simply a chart of the streamlines. For most practical purposes, the isobars may be used in place of the actual streamlines. The velocity field may be resolved into four separate components:

1. Translation,
2. Deformation,
3. Divergence,
4. Rotation.

The configuration of the streamlines is dependent on the relative magnitude of these various factors. Actually, sufficient data are rarely present to allow a velocity field to be analyzed into the several components mentioned above.

The *location* of the line of frontogenesis is determined solely by the distribution of the property (potential temperature) which is being affected by frontogenesis. Frontogenesis thus always occurs along a line where α is increasing most rapidly. Its *intensity* is governed by the angle between the line of frontogenesis and the *dilatation axis* of the velocity field.

DEFORMATION FIELDS (TWO DIMENSIONS)

A few relatively simple velocity fields, involving deformation only, will be considered here. These will give an idea of the method of locating the *dilatation* axis which is of great importance in estimating the intensity of frontogenesis. This discussion will include only two-dimensional fields. Actually all deformation fields in the atmosphere are three-dimensional, but for many purposes the two-dimensional treatment is entirely satisfactory, since the vertical component of motion is usually small. A later section will deal with three-dimensional fields, particularly as they affect the formation of inversions, and of convergent zones beneath fronts.

Pure *dilatation* is illustrated in figure 70-a, and pure *contraction* in figure 70-b. Neither of these conditions is likely to occur in

nature. They are useful, however, in explaining the effects of dilatation and contraction on a frontal zone. In both figures a frontal zone of definite width is shown before the deformation field acts (solid lines), and after it has acted for some time (dashed). The dilatation field spreads out the zone while the contraction field compresses it. The first case represents *frontolysis* and the second *frontogenesis*.

Whereas pure dilatation or pure contraction fields are unknown in nature, combinations of the two occur universally. Every synop-

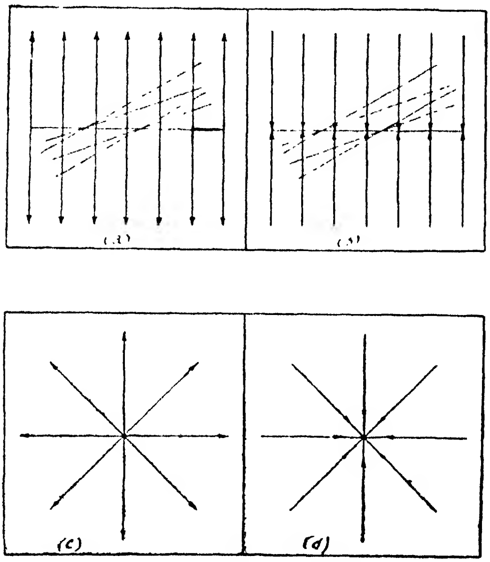

FIGURE 70.—(*a*) FIELD OF PURE DILATATION. (*b*) FIELD OF PURE CONTRACTION (*c*) FIELD OF PURE DIVERGENCE. (*d*) FIELD OF PURE CONVERGENCE

In (*a*) note spreading out of frontal zone in this field and in (*b*) note compressing of frontal zone.

tic weather chart represents one or more deformation fields showing dilatation and contraction and usually more or less ***divergence*** and ***rotation***. A pure deformation field is shown in figure 71-a. This results from the combination of the pure dilatation and contraction fields of figures 70-a and 70-b. It is unaffected by divergence (figure 70-c), or convergence (figure 70-d). ***Any frontal zone in this pure deformation field will rotate until it becomes parallel with the dilatation axis.*** (Actually it will only ***approach*** parallel-

ism, since an infinite time would be required to complete the rotation. Also, if the frontal zone is initially parallel to either the contraction axis or the dilatation axis, it will remain parallel to it.

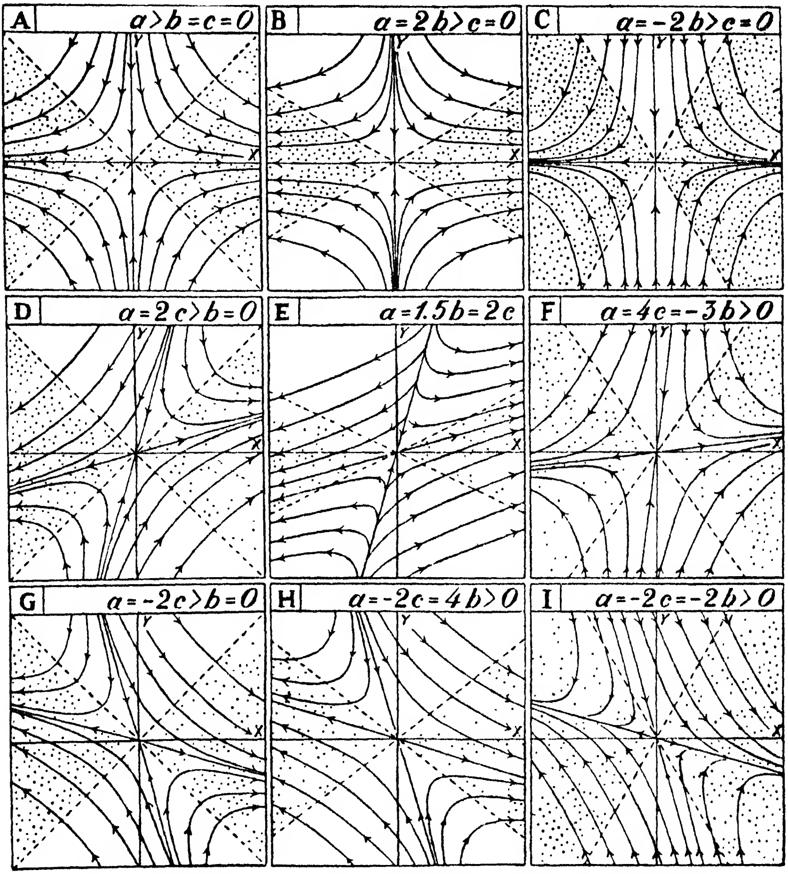

FIGURE 71.—TYPICAL HYPERBOLIC STREAMLINES

Sectors of frontogenesis are shaded; contraction axes, vertical; dilatation axes, horizontal, in all of the above figures. (a = deformation; b = divergence; c = rotation.) A = pure deformation. B = deformation + divergence. C = deformation + convergence. D = deformation + rotation. E = deformation + divergence + rotation. F = deformation + convergence + rotation. G, H, I, same as D, E, F, except with negative rotation instead of positive. (After Petterssen.)

These exceptions to the general rule are of no practical importance, however).

If a frontal zone that is nearly parallel to the contraction axis

(figure 71-a) starts rotating it will first be subject to frontolysis. During this interval the frontal zone becomes more diffuse. After it enters the shaded region, however, it is subject to frontogenesis and henceforth becomes more and more concentrated. Actually a considerable amount of time is required to effect the complete rotation of the frontal zone. For this reason, unless the frontal zone commences its rotation in the region of frontogenesis, it is unlikely that the deformation field will persist long enough to cause active frontogenesis.

CRITICAL ANGLE

The angle that the dashed line (figure 71-a) makes with the dilatation axis is called the *critical angle,* (ψ'). The larger the critical angle, the larger the *sector of frontogenesis,* and the greater the chance of a frontal zone being intensified. A few examples will serve to illustrate the practical use of these principles. Figure 72 represents the conditions along the east coast of Greenland. Here the line of frontogenesis (FF′) is near the coast, where the isotherms are most crowded. If the dilatation axis points northeastward, the frontal zone will be *parallel* to the dilatation axis and very active *frontogenesis* will occur. There will be no tendency toward rotation since the frontal zone is already parallel to the dilatation axis.

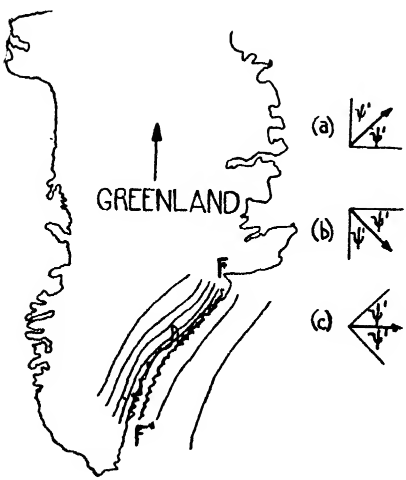

FIGURE 72.—FRONTOGENESIS ALONG THE EAST COAST OF GREENLAND

(*a*) Dilatation axis pointing *NE*; (*b*) dilatation axis pointing *SE*; (*c*) dilatation axis pointing *E*.

If the dilatation axis points southeastward, the frontal zone will be *perpendicular* to the dilatation axis (parallel to the contraction axis). There will be little tendency for rotation of the frontal zone and active *frontolysis* will occur. If the dilatation axis points

eastward, and the critical angle is 45°, the frontal zone will be rotated toward the dilatation axis, and immediately will be subjected to frontogenesis. The intensity of the frontogenesis will be small, however, until the frontal zone has rotated into closer parallelism with the dilatation axis. If the dilatation axis points slightly south of east, the frontal zone will be rotated toward it as before. First, it will be exposed to *frontolysis,* then after it comes within the critical angle, to *frontogenesis.* As mentioned above, however, unless the frontal zone is initially well within the critical angle it probably will never be subject to active frontogenesis because of the time required for its rotation.

STREAMLINE PATTERNS

The streamline pattern of figure 71-a is very simple and relatively rare. Here the lines of inflow and outflow are coincident with the contraction and dilatation axes. More common is the streamline situation illustrated in figure 71-d. Here a *rotational* component acts on the simple deformation field of figure 71-a. As a result, the lines of inflow and outflow (the straight streamlines) are no longer coincident with the axes. In such cases the axes may be located as follows:

1. Draw the lines that bisect the angle between the straight streamlines (lines of inflow and outflow).
2. Draw lines that lie 45° on either side of these bisectors.
3. The 45° line nearest the outflow streamline is the *dilatation axis* and the one nearest the inflow line is the *contraction axis.*

This construction may readily be sketched directly on the weather chart, using the actual streamlines or the isobars, to obtain the direction of the *dilatation axis.* Once this is obtained the chances for frontogenesis may at once be estimated when the location of the line of frontogenesis is known. This may, of course, be found from the temperature field. The importance of these considerations for practical forecasting can scarcely be overestimated. The forecaster can tell at a glance whether a pre-existing front or a general frontal zone will tend to increase or decrease in intensity. He can tell where the general circulation of the earth is most likely

to produce frontal zones and where these will be subjected to frontogenesis or frontolysis.

Many other types of streamline patterns occur in nature but those illustrated above include the more common forms. The evaluation of the critical angle is generally rather difficult, especially in cases involving convergent or rotational components of the velocity field. Generally it suffices to consider that it is roughly equal to 45°. When considerable convergence is present it will be larger, when divergence is present it will be smaller.

CYCLOGENESIS AND ANTICYCLOGENESIS

The streamline patterns discussed in the preceding section ordinarily do not persist for any length of time. Whenever definite divergence or convergence exists, the velocity fields represented by figures 71-b and 71-c respectively tend to be transformed more or less rapidly into anticyclonic or cyclonic circulations, respectively.

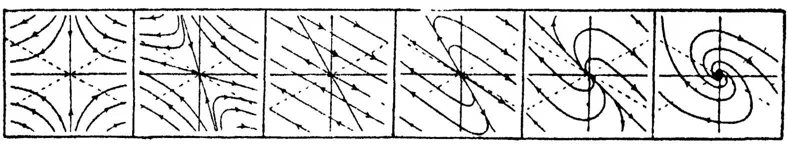

FIGURE 73.—FORMATION OF AN ANTICYCLONIC CENTER THROUGH DIVERGENCE

The third fourth and fifth patterns can persist for but a short time, and are essentially transitional. (After Petterssen.)

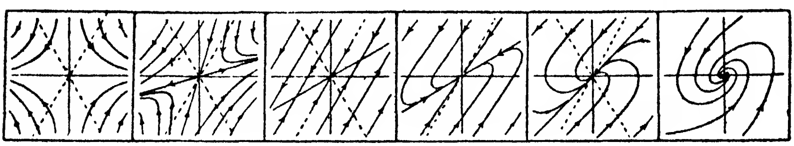

FIGURE 74.—FORMATION OF A CYCLONIC CENTER THROUGH CONVERGENCE

The third, fourth and fifth patterns can persist for but a short time, and are essentially transitional. (After Petterssen.)

This is illustrated by figures 73 and 74. In the first case divergence transforms the streamline field, by means of the steps indicated, into an anticyclone. The rotational effect within the anticyclone cannot exceed a certain definite value which depends on the latitude. This agrees well with actual weather chart situations. In the second

case convergence changes the streamlines, by means of the steps indicated, into a cyclone. Here there is no limit to the rotational effect.

It is very likely that tropical hurricanes often originate by the breakdown of a velocity field due to convergence in the manner suggested by figure 74. A relatively inactive deformation field in the tropics is thus often affected by convergence due to widespread convection over the ocean. This is probably sufficient to initiate cyclonic flow and result in the genesis of these tropical disturbances (see page 252).

It may be seen from the above discussion that cyclones may originate without pre-existing fronts. The only necessity for cyclogenesis is a convergent deformation field. Actually, most if not all extratropical cyclones originate as waves along frontal surfaces, but *tropical cyclones,* or hurricanes, are usually entirely unconnected with fronts.

DEFORMATION FIELDS (THREE DIMENSIONS)

In three-dimensional deformation fields, three deformation axes are present. Since the total *volume* of such a deformation field does not change appreciably the deformation axes cannot all have the same sign. The motion along one axis must thus be different from that along the other two and also be of greater magnitude in order to offset the effect of the others. This axis is termed the *principal axis.* The plane containing the *secondary axes* is termed the *deformation plane.* This is of course perpendicular to the principal axis.

Deformation in such a system may consist either of:

1. Dilatation along the principal axis, and contraction along the two secondary axes or,
2. Contraction along the principal axis, and dilatation along the two secondary axes.

In nature the principal axis is often vertical. Thus, when contraction occurs along it, dilatation occurs in the deformation plane, which in this case represents the surface. This type of activity is illustrated in figure 75. Here, isothermal surfaces tend to be concentrated near the surface and conditions are favorable for the

production of *inversions*. This situation commonly occurs in subsiding anticyclones.

If the principal axis is a horizontal contraction axis the deformation plane will be a vertical dilatation plane. This tends to destroy inversions and to intensify fronts. Figure 75 (below) shows the condition for frontogenesis, with a horizontal contraction axis

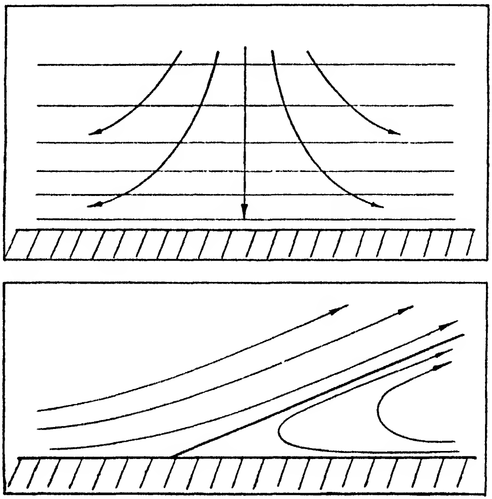

FIGURE 75.—(ABOVE) FORMATION OF INVERSIONS THROUGH SUBSIDENCE. (BELOW) FRONTOGENESIS THROUGH HORIZONTAL CONVERGENCE

In the above figure the principal axis of the deformation field is vertical, with dilatation in the deformation plane, in this case the surface. In the figure below the principal axis is a horizontal contraction axis. The deformation plane is sloping and is subject to dilatation.

and a sloping deformation plane showing dilatation. The convergent airflow in such situations leads to frontal activity. The convergence beneath the frontal surface often causes very important weather phenomena (page 311). When a front lies in a deformation field where the principal axis is a vertical contraction axis, the front will be subject to frontolysis—figure 75 (above).

SUMMARY

Several conditions must be fulfilled before a front may be formed.

1. The surfaces of constant potential temperature of the air masses must be sloping surfaces. If they are horizontal surfaces, the horizontal contraction which occurs with frontogenesis will not be able to concentrate

them. (This condition also states implicitly that the air masses must be stratified, and not have the same potential temperature throughout.)

2. The potential temperature surfaces must make a relatively small angle with the dilatation axis. If the angle is greater than the critical angle, frontolysis will occur.

3. The density difference at the same level must be such that a circulation tends to develop, which will cause the cold air to spread out and underlie the warm air. If the cold air already underlies the warm air, as in an inversion, there is no *potential energy of mass distribution* and no tendency toward frontogensis.

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CHAPTER 11

STRUCTURE OF EXTRATROPICAL CYCLONES

INTRODUCTION

It has been shown by V. Bjerknes, J. Bjerknes, Bergeron and others that most extratropical disturbances in the atmosphere are the result of wave action along discontinuity surfaces or fronts. V. Bjerknes brilliantly applied the principles of classical hydrodynamics to meteorology and clearly demonstrated the wave nature of extratropical cyclones. J. Bjerknes and T. Bergeron showed by means of numerous practical examples, the wide application of the pioneering work of V. Bjerknes. Their work has since been greatly refined by them and by other meteorologists.

An extratropical cyclone generally originates as a small irregularity in a pre-existing frontal surface. This rapidly increases in intensity to form an active wave disturbance. Since the density of the atmosphere is relatively low, the wave length of such disturbances is generally very large, commonly being of the order of from 500 to 1500 miles. The extratropical cyclone may exhibit an exceedingly wide variation in its structure. In the simplest cases it may merely be a slight wave along a frontal surface. Again it may involve several fronts. In the more complex situations, a number of fronts, both at the surface and aloft, may be involved. Besides the complexity within individual cyclones, they may be inter-connected in innumerable ways. It is thus hardly possible to present type examples that will represent all, or even most, of the situations encountered in nature. To facilitate study of the extratropical cyclone, however, it will be found convenient to discuss certain types which have been found to occur frequently.

THE STABLE WAVE CYCLONE

The simplest form of the extratropical cyclone is that of a *stable wave* along a discontinuity surface between two air masses

This cyclone consists of a cold front and a warm front, with an area of low atmospheric pressure at their junction (figure 76). In most instances the stable wave cyclone changes its structure very

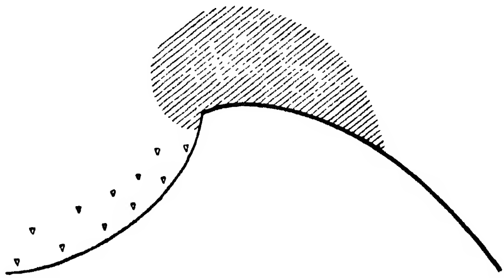

FIGURE 76.—STABLE WAVE CYCLONE SHOWING CHARACTERISTIC DISTRIBUTION, OF PRECIPITATION

Wide rain area appears in advance of warm front, and showery conditions occur along and to the rear of, cold front.

little during its life *unless occlusion occurs.* After occlusion the structure undergoes rapid modification as will be seen later.

As long as the cyclone continues as a stable wave, it deepens very little. Therefore, the appearance of strong tendencies in connection with this type of cyclone invariably heralds the onset of active occlusion. The wave generally moves along the front with a fairly constant velocity as long as it remains stable. The wave length, or distance between crests of the stable waves, is usually very constant during a particular series of waves. It is also relatively constant in a given region during any season.

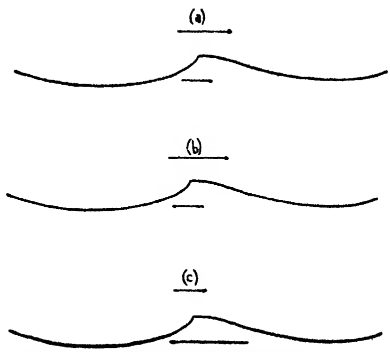

FIGURE 77.—CONVECTIVE COMPONENT OF VELOCITY OF A STABLE WAVE CYCLONE

In cases (*a*) and (*b*) the net velocity is toward the right, and the wave will move in this direction, provided the dynamic component is exceeded by the convective. In case (*c*) the net velocity is toward the left. Here the direction of motion depends on the balance between the dynamic and convective components.

Velocity—The velocity of a stable wave depends on two factors: (1) the convective component, and (2) the dynamic component. The convective component is approximately equal to the

average of the wind velocities on the two sides of the front. In figure 77 the convective component of the velocity is shown for three types of wind distribution. The first two cases yield a convective component directed to the right, and the last case a component toward the left. The dynamic component of the velocity is largely a result of the deflective force due to the earth's rotation (the Coriolis force). This causes air which is overrunning the warm front of a cyclone to be deflected so that it yields a dynamic component toward the right in the northern hemisphere and toward the left in the southern hemisphere. In most cases this produces a deflection *toward the east.*

The total propagating velocity of any stable wave is the net effect of the convective and dynamic velocity. Since the value of the dynamic component is usually directed toward the east, and since it varies but little in any one region, the convective component is the more important one in actual forecasting. If both of the components act in the same direction the wave will move rapidly, with a velocity equal to their sum. If they act in opposite directions the wave will move slowly, or even *retrograde* if the convective velocity exceeds the dynamic. This case occurs frequently in the lower latitudes. Here the dynamic velocity is small since it depends on the Coriolis force, and the convective velocity is often directed westward in the Trade Wind belt or in the Horse Latitudes.

Weather Conditions—Although the structure of the stable wave cyclone remains unchanged during its history, the weather phenomena associated with it may vary considerably as the air masses involved and the intensity of the circulation vary. These factors have a pronounced effect upon the cloud systems and precipitation areas associated with the movement of such storms and should be given careful consideration. In following the development of this type of disturbance the barometric tendencies should be observed carefully in order to detect the earliest indications of any tendency toward occlusion. As long as the disturbance remains a stable wave the tendency values involved will, in general, be rather small, so that the sudden appearance of any large tendencies is generally reliable indication of the beginning of occlusion.

In this type of storm the precipitation area usually exhibits a pattern similar to that in figure 76, with most of it of the warm front type. The extent of the warm front precipitation area depends

very largely on the structure of the warm air mass. With relatively dry air having a high condensation level and considerable stability, the precipitation area may be very small and uneven. If the overrunning warm air possesses a large amount of moisture and is fairly stable the precipitation area may be very large with rather uniform intensity throughout. If the warm air has a high humidity and is convectively unstable the precipitation area may be large with very irregular distribution.

Similarly, the precipitation occurring along the cold front is related chiefly to the air mass properties of the warm air. When the warm air is relatively dry no precipitation at all will occur. If it is rather moist, but stable, the precipitation area will be narrow and the precipitation will be of moderate intensity. The precipitation *within* the cold air mass depends entirely on the air mass properties of the cold air. If this is dry and stable there will be no precipitation at all, whereas, if it is fairly moist and unstable a number of scattered showers may be expected.

The importance of determining the rate of movement of stable waves as early as possible is well illustrated along the Atlantic coastal region. At times when a quasi-stationary front lies in a northeasterly-southwesterly direction along this region stable waves frequently travel northeastward along the front, causing large areas of unfavorable weather. These waves commonly develop during the spring and autumn months of the year during outbreaks of Polar Canadian air. As these outbreaks move southeastward over the United States they frequently lose intensity over the southern states and their forward movement ceases.

The early recognition of these waves is very important because of the large dynamic and convective components of their velocity. They are generally first indicated by the appearance of an *isallobaric low* along the front. A stable wave rapidly develops which is propagated northeastward along the front. The speed of propagation depends very largely on the wind velocities along the front in the intermediate levels, as the dynamic component due to the Coriolis force is fairly constant in this general region. Since these waves frequently move at velocities of 60 to 80 miles per hour along the front, causing rapidly moving precipitation areas, it is obvious why their early recognition is essential to weather forecasts for the middle and north Atlantic coast. Occasionally several such stable waves may occur along the stationary front at the same time, each

one with its associated unfavorable weather area. If the waves move rapidly the precipitation area may pass over a given region in a space of a few hours while, if they move slowly, the weather at any point may remain poor for a considerable length of time. This type of storm frequently originates at the southern extremity of the Appalachian Mountains. Here, there is a marked tendency for the cold air to protrude into the warm sector, thus giving the initial impetus necessary to cause the formation of a wave. (See page 85.)

Occasionally these waves form some distance offshore. Their early recognition in such cases is very difficult because of the lack of sufficient weather reports. Such waves may move northeastward along fronts lying just offshore producing a precipitation area which moves up the coast line. This type of activity may be expected whenever a cold front moves offshore with a very slight forward movement. At times when the fronts move offshore with a considerable velocity, they generally do not become subject to wave formation until they have passed quite a distance out to sea.

Waves which form in the general region between the Atlantic coast and Bermuda, often deepen and proceed eastward across the Atlantic Ocean as regenerated major disturbances. In fact, this general area is an active region of frontogenesis since the contrast in temperatures between the cold continental region to the northwest and the warm ocean surface to the south, together with the prevailing wind, tend to concentrate the solenoid field. The position of the Azores HIGH determines largely the movement of waves formed along fronts in the general region of the southeastern United States and the adjoining ocean. If the HIGH is centered well to the north so that the wind directions along the south Atlantic coast are generally southeasterly and of rather low velocities the waves will generally move comparatively slowly. If the HIGH is centered well to the south, however, so that the winds over the south Atlantic coast are southerly or southwesterly, the convective component along the front will be large and the waves will move rapidly.

THE UNSTABLE WAVE CYCLONE—THE OCCLUSION PROCESS

During the period while a cyclone remains a stable wave it retains a simple structure. As it propagates it may deepen slightly

and increase its amplitude somewhat but in general it remains unchanged. As soon as *occlusion* commences, however, it rapidly becomes unstable. From this point on, the wave may develop very rapidly and with great diversity. All of the meteorological elements —temperature, pressure, humidity, wind direction, etc.—as well as the type of surface and the topography affect its structure. It is in the interaction of all these various elements that is found an explanation for the great complexity of cyclonic storms. No two

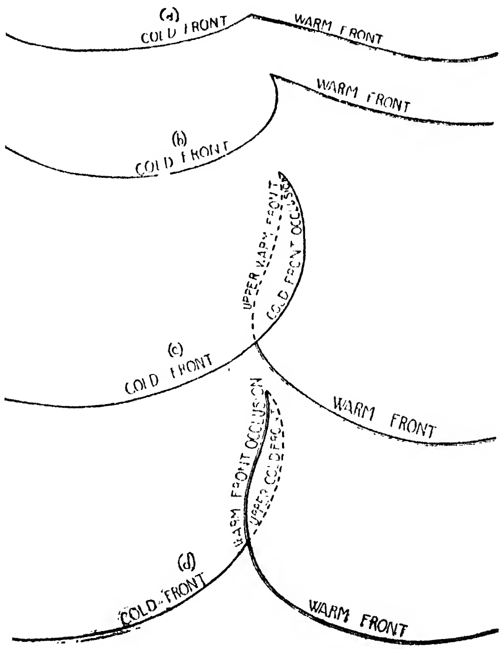

FIGURE 78.—OCCLUSION OF A STABLE WAVE CYCLONE

(*a*) The stable wave before occlusion; (*b*) beginning of occlusion; (*c*) cold front type occlusion partially occluded; (*d*) warm front type occlusion

are exactly alike. The development of two essentially similar situations may take place in very different manners. The forecasting of these developments with the many problems involved is one of the chief duties of the meteorologist.

The occlusion process involves the forcing aloft of a portion of the warm sector of a cyclone. In figure 78-a is shown the situation preceding the onset of occlusion. This figure represents a typical stable wave cyclone and little variation from this example occurs

while the cyclone remains stable. As break-down of the stable state approaches the fronts generally take on the configuration shown in figure 78-b. This is but a momentary state and actual occlusion follows at once. The occlusion may proceed as in figure 78-c, where the air in back of the cold front (*c-c-c*) is denser than either the air immediately in front of it or the air in front of the warm front. This is the cold front type occlusion. It may also proceed as in figure 78-d, where the air behind the cold front is denser than the air immediately in front of it, but lighter than the air in front of the warm front. This results in a warm front type occlusion.

In either case a portion of one of the fronts is forced aloft. In the cold front type occlusion the warm front near the center of the cyclone is forced aloft (figure 78-c). In the warm front type occlusion a portion of the cold front near the center is forced upward (figure 78-d). The upper warm front of the cold front type occlusion has little importance in the future development of the cyclone and is rarely indicated on synoptic charts. The upper cold front of the warm front type occlusion, however, is often of considerable importance and it should always be shown on the weather map.

The development of the unstable wave after occlusion has commenced depends upon:

(1) the rate of movement of the wave as a whole, and
(2) the rate of movement of the fronts within the wave.

If the crest of the wave (located at the center of low pressure) moves with nearly the same velocity as the cold front, the development of the cyclone will take place as shown in figure 79-a, b, c. This represents a rather typical development of a cold front type occlusion undergoing slow decay as it approaches a region of frontolysis.

If the crest of the wave is considerably retarded while the frontal system continues to move rapidly, the cyclone often develops as indicated in figure 79-d, e, f. Here the occlusion tends to become "rolled-up" due to intense circulation around the cyclonic center. Actually this situation is the closest approach seen in nature, on a large scale, to true dynamic instability which tends to reach the spiral form shown in figure 80. This extreme case of dynamic instability is probably reached on a smaller scale in the tornado.

The so-called "bent-back occlusion," of Bjerknes, illustrated in figure 81-b, may occur occasionally. This might be expected in cases where the cyclonic center is retarded as in figure 79-e. As a

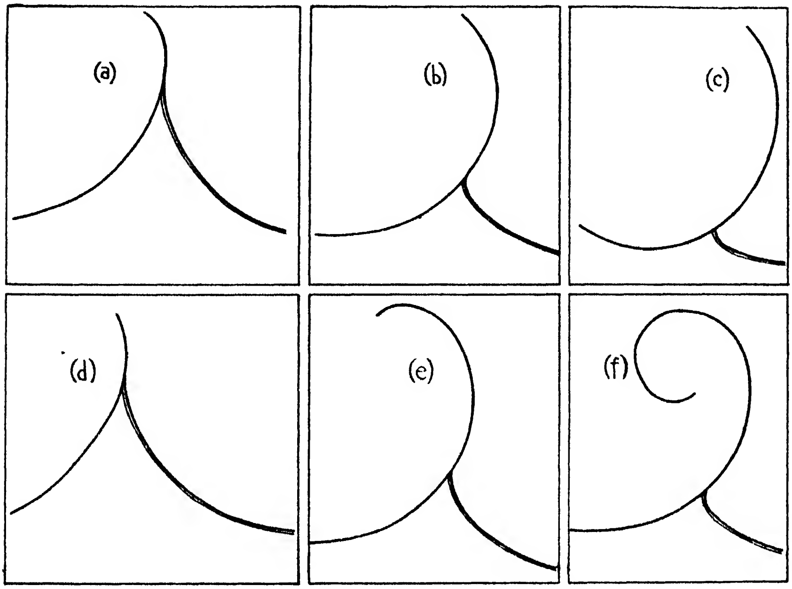

FIGURE 79.—DEVELOPMENT OF THE UNSTABLE WAVE CYCLONE

(*a*), (*b*), (*c*) represent cases where the crest of the wave moves with approximately the same velocity as the cold front. (*d*), (*e*), (*f*) represent cases where the crest is considerably retarded and the occlusion tends to become "rolled-up."

matter of fact most cases of "bent-back occlusions" turn out on careful examination to be secondary cold fronts. The two situations appear very similar superficially (figure 81), but it is generally only in regions where weather reports are scanty that the "bent-back occlusion" survives careful scrutiny!

Regenerated Cyclones—As decaying cyclonic storms pass into regions of cyclogenesis and frontogenesis they are very frequently regenerated to form active disturbances. In most cases this is caused by an influx of warm, moist air resulting from a gradual shift in the general circulation. Thus, well occluded storms which pass eastward over the Rocky Mountain region are often regenerated as the general circulation in the plains area carries Tropical Gulf air northward into the decaying cyclone. As a result of this activity a fresh wave may form on the old occlusion and deepen rapidly as

a major cyclone (figure 82). If the air which moves northward is a dry type with but little available energy of condensation, the cyclone will deepen but little, while if the air is moist, with a large amount of available energy, it may deepen rapidly. In any case the important synoptic event leading to the deepening process is the influx of a fresh supply of warm air.

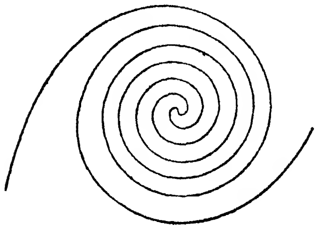

FIGURE 80.—DYNAMIC INSTABILITY

This state is probably approached in nature by the tornado, and to a less degree by well "rolled-up" occlusions.

Not infrequently, regeneration of an old storm occurs at the junction of the cold front with the occlusion (figure 83). This often occurs during the latter his-

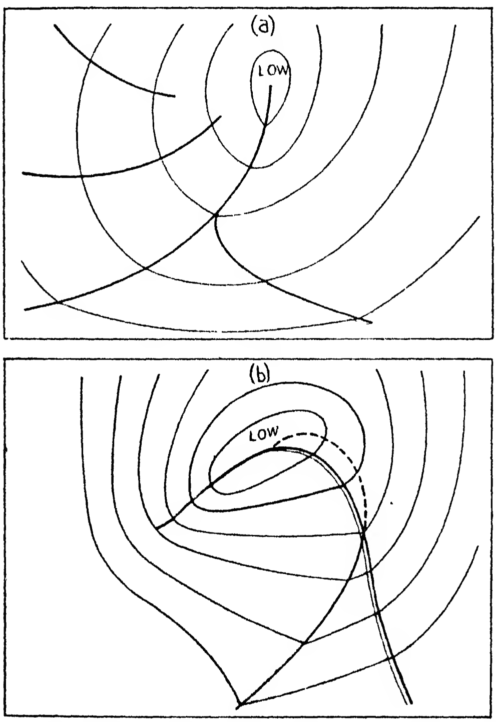

FIGURE 81.—(*a*) PARTIALLY OCCLUDED UNSTABLE WAVE CYCLONE WITH SECONDARY COLD FRONTS. (*b*) "BENT-BACK OCCLUSION"

Note the resemblance of these two types of structure. Most cases of "bent-back occlusions" prove to be secondary cold fronts when carefully analyzed.

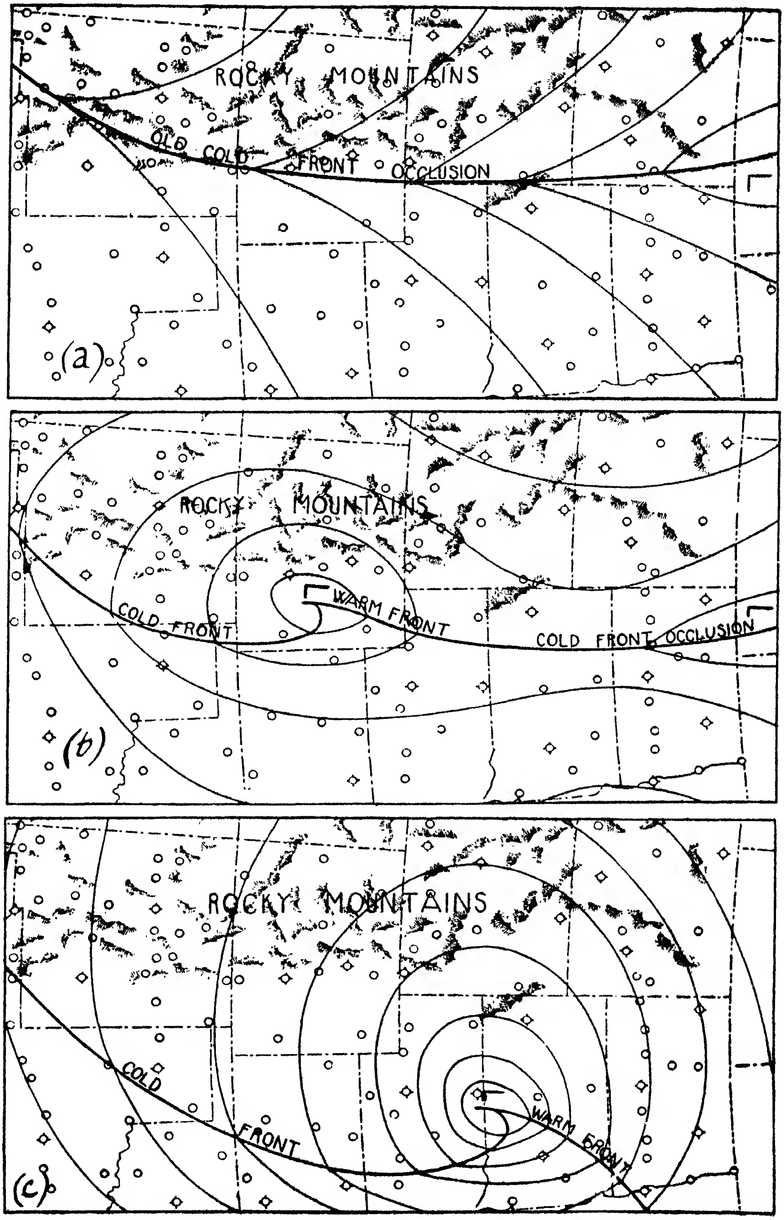

FIGURE 82.—REGENERATION OF OLD OCCLUSION EAST OF THE ROCKY MOUNTAINS BY INFLUX OF WARM AIR FROM THE SOUTH

(a) Old occlusion just after passing east over mountains. (b) Beginning of regeneration with formation of separate LOW center. (c) Rapid deepening of system.

tory of a cyclone when the lower portion has pushed into the trade wind belt. A supply of fresh tropical air becomes available at such times and the movement of the cyclone becomes very sluggish as the dynamic component of the velocity becomes slight and the convective component takes on an east to west motion. Secondary cyclones formed in this manner frequently persist for long periods causing extensive areas of unfavorable weather. They often occur during the cold season in the north Pacific Ocean between Hawaii and the mainland. At such times weather conditions may remain unfavorable for days at

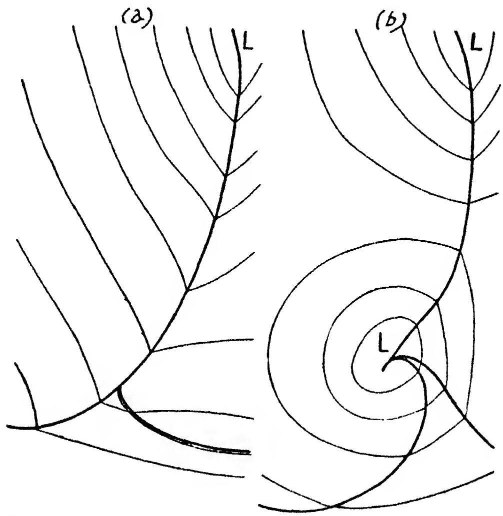

FIGURE 83.—REGENERATION OF OLD OCCLUSION AT JUNCTION OF WARM AND COLD FRONTS

a time as the storm passes through its slow development. Often the center will not shift its position more than a few miles per day due to the almost perfect balance between the dynamic and convective components of its velocity. If the center moves a few degrees to the north it may start a slow progress eastward; if it moves south it will retrograde westward. The typical *Kona storm* of the Hawaiian region is generally a result of the passage of such a regenerated cyclone in that general vicinity.

An interesting form of cyclonic development frequently occurs over the region just east of the Rocky Mountains as indicated in

figure 84. Here a fresh outbreak of Polar Pacific air has just crossed the Rocky Mountain region behind the cold front. Under such conditions a secondary cyclone very frequently arises along the northern portion of this occlusion due to the large temperature contrasts that exist under such cases. This cyclone commonly proceeds southward as a stable wave along the frontal surface as shown in the figure, generally becoming unstable at some point in southern

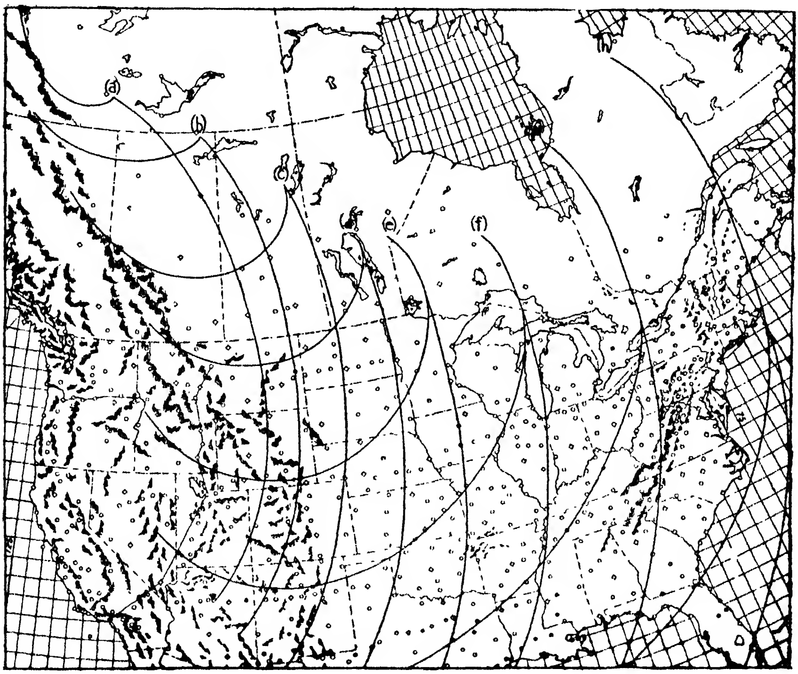

FIGURE 84.—FORMATION OF WAVE ON POLAR PACIFIC COLD FRONT AND SUBSEQUENT OUTBREAK OF POLAR CANADIAN AIR WITH DEVELOPMENT OF UNSTABLE WAVE

Canada and then proceeding southeastward as an unstable wave. This is such a common phenomenon in connection with outbreaks of Polar Pacific air during the winter that it may be said to be characteristic of this situation. In fact, many of the major outbreaks of Polar Canadian air over southeastern Canada and the central and eastern United States during the winter are brought about in this manner. Apparently the eastward movement of a Polar Pacific current into the Polar Canadian source region with the

production of a small secondary cyclone provides the necessary impetus to cause a breakdown of the Canadian anticyclone and the occurrence of a cold outbreak. It will be noted that almost all of the major cold waves are preceded by the appearance of Polar Pacific air behind a cold front occlusion. The recognition of this type of situation is thus of considerable importance to forecasting in the middle west and eastern portions of both Canada and the United States.

This type of situation may be expected to occur in any region where the source areas of continental air are invaded by relatively warm maritime air in the winter season. Thus, over central and eastern Europe the production of outbreaks of Polar Continental air may be expected when polar maritime air from the North Atlantic passes inland over the great reservoir of Polar Continental air of eastern Europe and Siberia.

SECONDARY FRONTS

Certain synoptic situations are particularly favorable to the formation of secondary cold fronts. The essential features for this development are, the rapid movement of a mass of cold air, and a strong latitudinal temperature gradient. When these conditions are fulfilled, as during a strong outbreak of Polar Continental air during the winter, secondary cold fronts are almost certain to develop. The secondary fronts may be expected to form as long as the wind velocities in the cold air mass continue high. As the air flow decreases the latitudinal temperature gradient diminishes and the formation of secondaries ceases.

Figure 85 illustrates the formation of secondary cold fronts during a strong outbreak of Polar Canadian air over the eastern part of the United States. The secondaries are commonly separated by from 200 to 400 miles and move with velocities of from 25 to 40 miles per hour. Thus, a continuous succession of them may pass a given locality at intervals of 6–12 hours. At times the secondary fronts represent merely a slight concentration of the postfrontal instability, while again they may represent cold front activity of nearly the intensity of that accompanying the original cold front.

When the major cold front of a strong outbreak of cold air is retarded and subjected to wave formation the secondaries will almost invariably show the same features. This peculiarity of sec-

ondary fronts is illustrated in figure 85. It will be noted here that the incipient wave in the main front at *A*, is duplicated in the secondary front at *B*. As the wave in the main front develops and

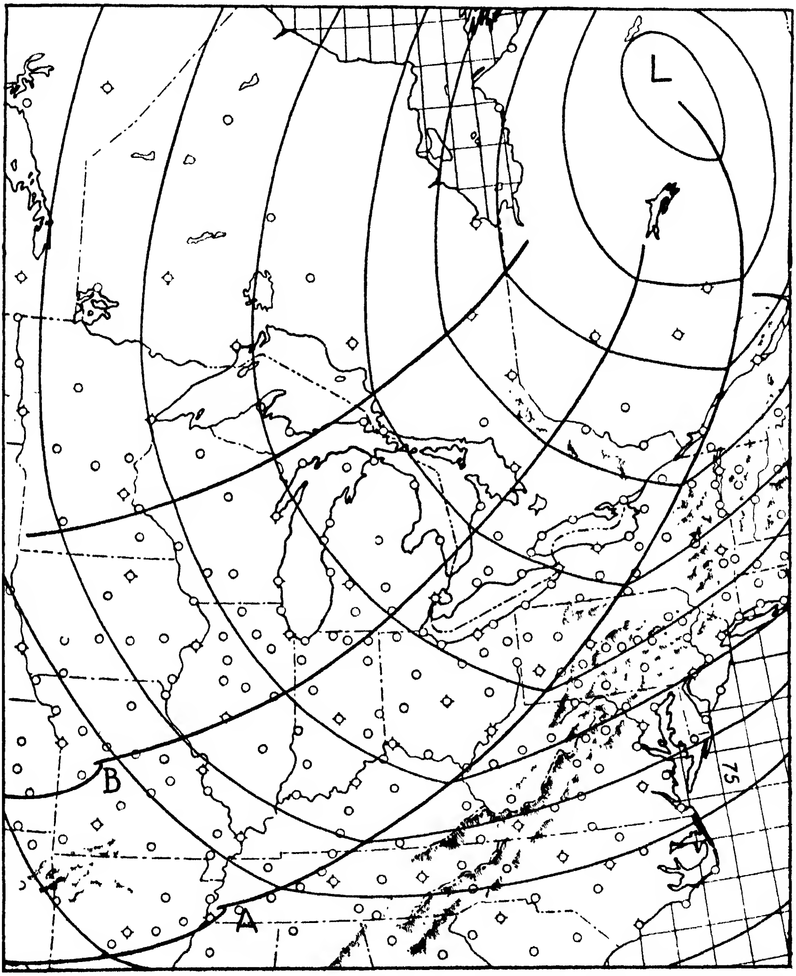

FIGURE 85.—FORMATION OF SECONDARY COLD FRONTS IN POLAR CANADIAN AIR IN THE EASTERN UNITED STATES

Note tendency for repetition of minor features in adjacent fronts as at *A* and *B*.

moves along the front the wave at *B* also moves in a similar manner. Occasionally the same feature may be repeated in two or more secondaries.

MODIFICATION OF CYCLONES DURING INTENSE CIRCULATION

The intense circulation which accompanies deep cyclonic centers generally effects rapid modification of their structure. Complicated frontal systems are broken down into simple forms by the mixing of the air masses. Particularly is this true of regenerated old cyclones, in which a complex system of surface and upper fronts may be transformed within a short time into a relatively simple frontal system. It is thus frequently observed that relatively weak and complex cyclonic systems over the Rocky Mountain region are regenerated as they reach the western Great Plains due to the influx of warm air from the Gulf of Mexico. The initially complex structure disappears and is replaced by a relatively simple system of fronts. This is accompanied frequently by rapid deepening of the system together with rapid movement of the fronts. In such cases the fronts tend to rotate around about the center of the cyclone as spokes around a wheel.

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This is mainly a review of the more recent developments in dynamic meteorology. No attempt is made to add any new ideas.

2. Discussion and Illustration of Problems Suggested by the Analysis of Atmospheric Cross-Sections: H. C. Willett. Mass. Inst. Tech. Papers in Phys. Oceanog. and Meteor., v. 4, n. 2, 1935.

Willett presents some of the more important problems which occur to anyone constructing atmospheric cross-sections, including the origin of various types of tropical air, the location of upper fronts, etc.

3. Atmospheric Structure over the Southern United States, Dec. 30-31, 1927. Determined with the Aid of Sounding Balloon Observations: Gardner Emmons. Mass. Inst. Tech. Meteor. Course Prof. Notes, n. 9, 1935.

This paper presents a detailed study of atmospheric structure during an interesting synoptic situation, in which aerographic soundings gave a rather complete picture of the structure in three dimensions.

4. Investigations of Selected European Cyclones by means of Serial Ascents: J. Bjerknes. Geof. Publik. v. 11, n. 4, 1935.

This presents further examples of the use of meteorograph soundings in determining storm structure. Bjerknes presents the atmospheric cross-section as a useful means of studying storm structure.

5. Exploration de quelques perturbations atmosphériques à l'aide de sondages rapprochés dans le temps: J. Bjerknes. Geof. Publik. v. 9, n. 9, 1932.

This is a most interesting paper in which Bjerknes employs meteorograph soundings over a wide area to determine the structure of a storm, and to check his analysis prepared from surface observations. In French.

6. The Extra-Tropical Cyclones of Eastern China and Their Characteristics: S.-W. Sung. Mem. Natl. Res. Inst. Meteor. Nanking, 1931.

This gives a brief summary of some of the more important types of storms of eastern China.

7. Über die dreidimensional verknüpfende Wetteranalyse: T. Bergeron. Geof. Publik., v. 5, n. 6, 1928.

This very important paper, all in German, presents Bergeron's work on the deformation field, with practical applications, in addition to much new information concerning the mechanics of cyclones. He gives a number of examples of practical application of his ideas.

8. Wellen und Wirbel an einer quasistationären Grenzfläche über Europa: T. Bergeron and G. Swaboda. Veröffen. d. Geophy. Inst. d. Univ. Leipzig., v. 5, n. 4, 1924.

This presents a practical analysis of an interesting synoptic situation. It shows the presence of both stable and unstable waves along a front which extended across Europe. All in German.

9. Life Cycle of Cyclones and the Polar Front Theory of Atmospheric Circulation: J. Bjerknes and H. Solberg. Geof. Publik., v. 3, n. 1, 1923.

This paper deals with the larger aspects of storm structure, including cyclone families and interzonal circulation over the earth.

10. On the Dynamics of the Circular Vortex with Applications to the Atmosphere and Atmospheric Vortex and Wave Motions: V. Bjerknes. Geof. Publik., v. 2, n. 4, 1921.

This is a very important technical paper presenting the work of V. Bjerknes on classical hydrodynamics as applied to meteorology. He gives the theory of discontinuity surfaces and waves along them.

11. Meteorological Conditions for the Formation of Rain: J. Bjerknes and H. Solberg. Geof. Publik., v. 2, n. 3, 1921.

This paper elaborates considerably on the first paper, mentioned below. It explains clearly the cyclone model as conceived by the Norwegian meteorologists.

12. On the Structure of Moving Cyclones: J. Bjerknes. Geof. Publik., v. 1, n. 2, 1919.

This is the first paper in the classical series of papers issued by the meteorologists of the Norwegian weather service dealing with the practical applications of classical hydrodynamics to weather forecasting. It is a short paper, barely outlining the main features of cyclonic structure.

CHAPTER 12

THUNDERSTORMS—TORNADOES—HURRICANES

INTRODUCTION

There are three principal types of thunderstorms, each of which may be readily recognized from the synoptic situation:

(a) air mass thunderstorms,
(b) cold front thunderstorms,
(c) warm front thunderstorms.

The characteristics of each of these differ somewhat from the others, especially in regard to the elevation at which maximum turbulence may be encountered. At times these various types may occur together, or the air mass type may be intensified by frontal activity.

Thunderstorms present four chief sources of danger to airplanes: turbulence, hail, icing and lightning. Turbulence, when moderate, causes considerable passenger discomfort. When it is severe it may cause structural strain in an airplane. It may be avoided at times by choosing the altitude of minimum turbulence as determined from aerographic soundings. In regions of rough topography there may be no altitude of comparative quiet air, and here severe thunderstorms should be completely avoided. Hail only occurs in severe storms with strong vertical currents. It may cause structural damage when the hail stones are over ¼–½ inch in diameter. In known hail conditions thunderstorms should be avoided. Icing is usually very sporadic in occurrence in thunderstorms. Only rarely does it persist for a long enough time to constitute a definite hazard. Lightning is not apparently a hazard to modern, all metal, well bonded airplanes. It may give rise to very severe static however, and may even damage radio equipment sufficiently to render it inoperative. Severe lightning should be avoided since it is usually accompanied by strong turbulence and often, by hail.

AIR MASS THUNDERSTORMS

Air mass thunderstorms are those which occur well within air masses, unaffected by frontal activity. They are initiated by convective currents caused by surface heating. Conditionally unstable air is a necessity for their occurrence. They are common in the temperate zone only during the warm season. They may occur at any time of year in the tropics, but are much more frequent during the rainy season. They are very rare above latitude 45°, where unstable air and intense surface heating are usually lacking.

The energy of air mass thunderstorms is partly due to surface convective currents and partly to energy released during the condensation process. The Refsdal chart, or *emagram*, may best be utilized to show the processes which occur during the production of this type of thunderstorm. In figure 86, the curve *ABCDEFG* represents an aerographic sounding obtained during the early morning in the subtropics during the thunderstorm season. A surface inversion is present as shown by the position of the curve from *A* to *B*. This inversion is almost invariably present in soundings made early in the morning. The lower portion of the air, from the point *B* to the point *D*, is conditionally unstable to a marked degree with the lapse rate approaching the dry adiabatic. (For a discussion of conditional instability see pages 54-55.)

From point *D* up to point *E* the air is still conditionally unstable but to a much less marked degree than before. The lapse-rate here is still intermediate between the saturated and the dry adiabatic. Above *E* the air is absolutely stable, for both the saturated and the dry states. The stability still further increases above *F*.

As heating progresses during the day several changes will occur in the lower portions of the air mass. The upper portion, above *C* or *D*, will change but little. First, the surface will become heated and point *A* will move to the right along the line *AP*, representing the surface pressure. As heating continues, the lapse rate in the lower layers will become steeper and steeper until it reaches the dry adiabatic. Active mixing of the lower levels will then occur due to the absolute instability that results.

Convective Condensation Level—Mixing and convective lifting of these lower layers will continue more and more vigorously as the surface continues to be heated during the day. A point will

finally be reached when the ascending air columns will be cooled sufficiently to reach saturation. The level at which this occurs, the *convective condensation level* (CCL), is very important. It appears in figure 86, and represents the level of the bases of the cumulus clouds that will be formed.

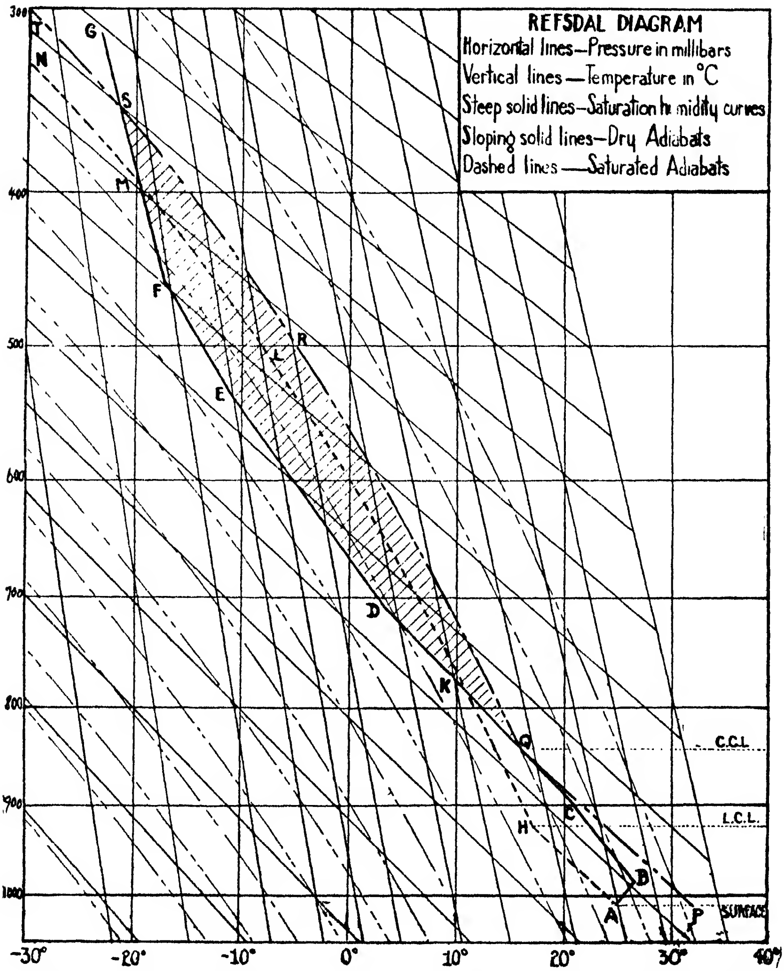

Figure 86.—Use of Refsdal Diagram in Forecasting Thunderstorms

All air movements below the CCL are concerned with the unsaturated or *dry* state. Air particles in this region therefore follow the dry adiabatic as they rise or fall. Above the CCL, rising air

particles follow the saturated adiabats. The CCL may be determined from a study of the sounding in the early morning. It lies at the intersection of the air mass curve (*ABCDEFG*), with the saturated specific humidity line representing the average conditions in the lower levels of the atmosphere. This may be determined by averaging the values of *W* for the lower levels (at points *A*, *B* and *C* in figure 86). Thus the point *Q*, in the figure lies at the intersection of the air mass curve, with the 14 gram saturation humidity line, and represents the elevation of the CCL in this case.

After a rising particle becomes saturated, at the CCL, it will continue to rise, now following the saturated adiabat (*QRST*). In the entire region from *Q* to *S*, the rising particle is *warmer* than its surroundings as represented by the portion of the air mass curve, *QDEFS*. Here, it will therefore tend to rise of its own accord since it is less dense than its environment. Above the point *S* it finally becomes cooler than its surroundings and tends to sink, since it is denser than its environment.

At all times that a rising particle follows the path *QRS* it is constantly being subjected to vertical accelerating forces due to the density differences. These forces reach their maximum at *R*, where the air mass curve is farthest removed from the saturated adiabat which the particle is following. Above *R*, the magnitude of the accelerating force decreases, although it still is directed upward. For this reason the vertical *velocity* of a particle continues to increase until *S* is reached. Above *S*, the accelerating force is directed downward and the vertical velocity rapidly decreases and becomes zero.

The size of the area between the air mass curve, *QDEFS*, and the path of the particle *QRS*, is of utmost importance in forecasting the occurrence and intensity of thunderstorms. If this area is small, the energy available for convection will be small and the thunderstorm will be of little importance. If it is large the storm will be intense. Note that this so-called *positive area* occurs only when the air mass curve lies *below* the curve representing the path of the particle. When the air mass curve lies above the particle curve the area is *negative* (*GST* in figure 86), and tends to *retard* convection.

It should be noted from the above discussion that air mass thunderstorms are connected with *conditional* instability. Only rarely do conditions within the atmosphere attain *absolute* instability,

and then in the lowermost portions only. Thus, to assume that thunderstorm activity may be expected only with an absolutely unstable lapse rate (steeper than the dry adiabatic) is entirely unfounded.

Lifting Condensation Level—A method of analysis which neglects the modification of the lower portions of the air mass due to convection during the day has been described by Refsdal. He assumed that a surface particle (*A* in figure 86) would rise along the dry adiabat through it until it became saturated at *H* (the *lifting condensation level*), then would follow the saturated adiabat through *H* (*HKLMN*). His analysis of the upper portions of the thunderstorm is the same as described above. His neglect of the modification of the surface layers due to surface heating, however, makes his analysis unreliable. He considers only the area *KLMFEDK* as positive area and the *ABCKHA* as negative area. Actually, the positive area should be much larger; *QRSMFEDKQ*, using the convective condensation level.

Choice of Flying Levels—The practical importance of the facts pointed out above, concerning rising particles in conditional unstable air, is considerable. Below the CCL the only upward impetus which rising particles have is that due to surface heating. Here, therefore, turbulence will not be especially intense. In this region downdrafts from cumulus clouds above may be encountered. These currents are caused by descending cold air from the interior of thunderstorms.

Above the CCL the vertical *accelerations* and *velocities* both increase steadily up to *R*, and the vertical *velocities* continue to increase up to *S*. The most turbulent region in a thunderstorm may be expected near *R*, where the vertical velocities are high and the accelerations reach their maximum. This level is commonly at about 12,000–15,000 feet above sea level. In mountainous regions the maximum turbulence is generally encountered higher—at from 18,000–20,000 feet above sea level. These figures have been borne out by a large number of actual observations and represent the experience of many pilots in flying through various types of thunderstorms.

This region of strong turbulence is obviously the worst level to attempt to fly through a thunderstorm. The best flying level to choose in a thunderstorm region is below or above the storm when possible. When the elevation of the terrain or the vertical develop-

ment of the storm makes this impossible, the lower levels of the clouds are comparatively favorable, since vertical currents are comparatively mild here.

Hail—When much of the positive area is above the freezing level as in figure 86, the chances for hail are excellent. The strong vertical currents generated in the lower levels carry water droplets upward until they freeze. These ice pellets tend to combine to form larger and larger hail stones until the vertical currents can no longer support them. They then fall out and reach the surface as hail stones, unless they melt in falling through the warm surface layers.

Severe hail storms may result from the intensification of air mass thunderstorms by frontal activity. Here, the extra lift imparted by the frontal movement causes increased vertical currents which can support large hail stones. Cold front thunderstorms and, in rare cases, warm front thunderstorms may also cause hail. In the tropics, where large amounts of convective energy are available, severe hail storms may occur without frontal activity, in thunderstorms that are purely convective in nature.

Time of Cumulus Formation—When the positive area lies largely below the freezing level no hail need be expected, and in fact the general cumulus development will be slight. When the positive area begins at a rather high elevation the cumulus activity will not commence until late in the day. When the CCL is near the surface, cumulus clouds will begin to form during the middle forenoon and thunderstorm activity may be expected by the early afternoon.

The time at which cumulus clouds will first form may be determined from a study of the emagram (figure 86). Convection due to surface heating will gradually mix the surface layers until a dry adiabatic lapse-rate is established up to the convective condensation level. (*PQ* in figure 86). At the moment the surface temperature reaches *P*, the condensation point will be reached by a particle at *Q*, on the dry adiabat through *P*, and condensation will commence. Thus, if the time at which the surface temperature will reach *P* can be determined, the time of the onset of cumulus activity may be determined rather accurately. This may be done by preparing a set of diurnal temperature curves for various stations at various times of the year, under different conditions of soil and sky conditions. Thus, if the diurnal curve for August at the station

represented by figure 86, with a moist soil (due to previous rains) and a clear sky, showed that the temperature *P* would be reached at 11 A.M. the forecaster could confidently predict the formation of cumulus clouds at about that time. The onset of thunderstorm activity will, of course, be delayed several hours since a very definite period of time is required for the rising particles to produce large thunderheads.

Sinuosites in the Pressure Field—The occurrence of thunderstorms in regions of relatively flat barometric gradient is frequently accompanied by small sinuosities in the isobars which are apparently unrelated to the main isobaric field. These slight irregularities in the isobars are generally found only on weather charts drawn during the daytime when convective activity is well marked. They evidently represent very minor cyclonic centers or troughs which accompany individual thunderstorms. In many cases they are apparently real sinuosities and not merely irregularities due to unreliable pressure reports.

When these fluctuations in the barometric field occur, it is generally advisable to construct isobars for each millibar in order to discover if the irregularities in the isobars are real or whether they merely represent accidental variations in the pressure reports. If they are real, they should appear at several adjoining weather reporting stations. The occurrence of these "micro-cyclones" is in many cases a reliable indication of the localities in which major thunderstorms will occur in a general situation which favors the development of air mass thunderstorms.

Topographic Influences—It has been observed frequently in relatively flat country, that certain regions have a much higher incidence of thunderstorms than others. This is particularly true in the Mississippi Valley region of the United States. In almost all cases these "breeding grounds" of thunderstorms are found to have a slightly higher elevation than the surrounding territory. The difference in elevation may amount to only a few hundred feet but apparently this slight difference produces sufficient additional lift, in convectively unstable air blowing across it, to act as a "trigger force" in setting off thunderstorms. This phenomenon is well marked in the region of southern Michigan. Here the low hills in the vicinity of Jackson, whose elevation above the surrounding country amounts to only 300–400 feet, show a very marked tendency to be the first to be affected by thunderstorms in that region. The same phe-

nomenon is noted at many other localities in flat regions. Thus the Ozark Mountains of the southern Missouri and Arkansas region, although they rise above the surrounding plain only 800 to 1000 feet, nevertheless show a much higher incidence of thunderstorms than the general surrounding area. Similarly in east Texas it is found that most of the thunderstorms which are observed, generally first appear in low rolling hills, such as those which lie a few miles north of Dallas, and those between Fort Worth and Eastland.

This tendency of thunderstorms to originate over areas which are slightly higher than the surrounding territory, provides a very useful means of detecting the first appearance of thunderstorm activity in a relatively uniform wide area, at any point of which thunderstorm activity may occur. The forecaster will find it well worth while, in relatively flat areas, to study carefully the detailed topography and to determine the points at which thunderstorms generally originate. These regions should then be watched closely during times of convective activity.

Thunderstorm Paths—In forecasting the movement of thunderstorms after they have once formed it is important to know the direction of the upper winds at levels from 8000 to 14,000 feet since it is found, almost invariably, that the thunderstorms move with the upper winds at these levels. This is particularly true for air mass thunderstorms whose direction is not related to frontal activity. As soon as the forecaster has noted the appearance of a well developed thunderstorm, he should at once estimate its path and rate of movement from the upper winds and then follow its movement across the country as indicated by successive reports from stations which it passes. This is very important for aviation purposes. A well developed thunderstorm may provide a very distinct hazard to flying and a knowledge of its path and rate of movement is essential. It is recommended that a large scale map of the region for which forecasts are issued be provided and that it be used for the express purpose of plotting thunderstorm movements. On this map the various weather reporting stations should be indicated and means should be provided for following the thunderstorm tracks by the use of colored pins.

Forecasts of thunderstorm movements made in this manner are extremely useful to airline operations, particularly as regards weather conditions at terminals. Arrangements may be made on the basis of thunderstorm forecasts not only to suspend operations in and out

of terminals at the time thunderstorms are expected but also to make necessary preparations to withstand the strong winds that frequently accompany such thunderstorms.

Thunderstorms which accompany the passage of cold fronts also move with the upper winds and frequently at a somewhat higher velocity than the cold front itself. Thus their path may cause them to appear some distance in advance of the actual cold front, especially at times when the front is being retarded. Warm front thunderstorms also follow closely the upper wind directions above the warm front surface and their paths may generally be forecast with considerable accuracy.

In following the paths of thunderstorms it is important to determine the approximate size of the storm area so that an estimate may be made of the time required for it to pass over a given point. This information can generally be obtained from the hourly and special weather reports issued at stations along its path. It is very useful in indicating the length of time that weather may remain unflyable at terminals which the storm's path crosses.

Aerographic Data—The forecasting of all types of thunderstorms depends primarily upon the use of upper air data obtained from aerographic soundings. Without this information it is impossible to know the temperature and moisture distribution in the upper atmosphere and thus it is quite impossible to know anything about stability conditions. In seasons of the year in which thunderstorms may be expected, therefore, it is very important for the forecaster to have access to this upper air information so that he may use it in preparing forecasts. It is, of course, possible to know in a general way the stability conditions in a given situation by a knowledge of the average air mass properties. This is of little value, however, in doubtful cases where the stability of the air masses involved is comparatively great. The practical value of aerographic information, from the standpoint of routine forecasting, during the thunderstorm season is therefore considerable. It is also important that aerographic soundings be carried to as high an elevation as possible, since the determination of the freezing level with relation to the regions of atmospheric instability is rather important. It has been found that if the freezing level lies at an elevation well above the region of maximum instability, the intensity of the thunderstorm will be rather small and that any hailstones formed will be insignificant.

COLD FRONT THUNDERSTORMS

In general the conditions outlined above for air mass thunderstorms apply also to those occurring along a cold front. The essential difference is the greater violence which is frequently observed in connection with cold front activity. Cold front thunderstorms are not at all dependent on surface heating since the lift necessary to release the conditional instability is provided by mechanical lifting above the cold front. So, this type of thunderstorm may occur at any time of day or night, whereas the air mass thunderstorm generally occurs at times of maximum contrast in temperature between the surface of the ground and the air:—in the afternoon over the land, and at night over the ocean. Furthermore, cold front thunderstorms may occur at almost any season of the year if the air mass which is being displaced is conditionally unstable. Of course, this instability is most pronounced in the warmer seasons of the year and may be entirely lacking during the winter, nevertheless, cold front thunderstorms frequently occur much earlier and later in the warm season than the air mass type. If a cold front passes over a region which is already affected by air mass thunderstorms the resulting increase in activity may be very pronounced.

It may be stated as a general rule that flying through a well developed cold front thunderstorm is always a hazardous undertaking, particularly if the air which is being lifted in front of the cold wedge shows marked instability. Under such conditions the vertical currents encountered not only may be extremely violent but they may also cause the formation of very large hailstones. With cold front thunderstorms the elevation of the saturation level can be determined as in the case of air mass thunderstorms by the moisture and temperature distribution within the lower atmosphere. The tops of cumulus clouds during strong cold front activity frequently reach very high levels, at times occupying the entire troposphere. In fact, the experience of pilots on recent high altitude flights indicates that cumulus clouds, even during the winter when the instability of the air masses involved is not excessive, may frequently reach altitudes in excess of 35,000 to 40,000 feet, so that flying over such disturbances is generally quite out of the question. Of course, the region of maximum vertical velocities never lies at such excessive altitudes and the vertical velocities encountered above 25,000 feet generally will be comparatively slight. On the other

hand the turbulence at low levels, even below the condensation level, may be rather severe in cold front disturbances, so there is no region of comparatively smooth air in an active cold front thunderstorm.

The general considerations with regard to cold front thunderstorms mentioned above apply also to cold front occlusions, although the activity along the occlusions is generally somewhat less unless the air in the warm sector, which has been forced aloft, happens to be especially unstable. Each case of this type demands separate consideration. The presence of an upper warm front may, in some cases, retard convective activity if the air above it is especially warm and dry, by providing an inversion level which cannot be penetrated by the air near the surface. On the other hand it may actually increase the general activity if the upper air is unstable itself.

WARM FRONT THUNDERSTORMS

A somewhat different type of thunderstorm occurs in connection with warm front activity when the overrunning warm air is conditionally unstable. This type of thunderstorm generally makes its appearance several hundred miles in advance of the surface position of the warm front. Generally a number of these storms form a line roughly parallel with the surface position of the warm front. These storms may occur at any time of day since they are not all dependent upon surface heating. They obtain the mechanical lift necessary to release the available conditional instability by lifting along the warm front surface. Frequently this type of thunderstorm makes its appearance rather suddenly, and at times unexpectedly, as the overrunning warm air reaches saturation. The occurrence of warm front thunderstorms usually can be forecast with considerable assurance: whenever instability within the warm air current is indicated by aerographic soundings, and wherever active overrunning is occurring. The actual point at which this type of thunderstorm may occur depends upon the slope of the warm front surface and the degree of saturation of the warm air and can be determined approximately if these quantities are known.

Figure 87 illustrates the general conditions under which this type of thunderstorm may occur. It is seen that the region of maximum turbulence is located considerably higher than in the case of air mass or cold front thunderstorms. Ordinarily flying conditions

beneath the warm front surface will be comparatively satisfactory. In general the warm front surface, at localities affected by this type of activity, will lie at elevations of from 4000 to 10,000 feet above sea level. Over flat country, therefore, it is a comparatively simple matter to avoid turbulence by flying beneath the thunderstorm

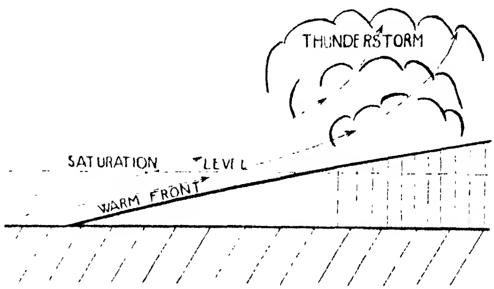

FIGURE 87.—FORMATION OF WARM FRONT THUNDERSTORM

This type of storm appears some distance in advance of the surface position of the warm front near the point where the overrunning warm air becomes saturated.

cloud. In rugged country the base of the clouds may approach the tops of the higher peaks and if these exceed 5000 to 8000 feet in elevation the warm front thunderstorm will have all the characteristics of a cold front disturbance. Generally speaking, the activity in a warm front thunderstorm is approximately that of a well developed air mass thunderstorm. Rarely will it approach the severity of the cold front type.

Thunderstorms produced in connection with warm front type occlusions are generally very similar to the ordinary warm front thunderstorms although they may occasionally be rather severe along the line marking the passage of the upper cold front as shown in figure 65. Here the lifting of the warm air passing up the warm front occlusion is considerably accelerated by the movement of the upper cold front. Since the air behind the upper cold front is generally comparatively dry, thunderstorm activity will almost invariably cease a short time after the passage of the upper front. As a result, in situations of this type, the thunderstorms will occur as a well defined band. This moves slowly across a region with the passage of the upper cold front. The onset of this type of thunderstorm is frequently very sudden, and unless care is taken to detect the earliest stages of the warm front occlusion process, this type of thunderstorm may appear quite unexpectedly.

PRECIPITATION IN THUNDERSTORMS

It has been shown that in most cases, rain does not begin to fall until the rising air currents have ascended above the freezing level. This is due, according to Bergeron's theory (see chapter 13) to the breakdown of colloidal stability with both the liquid and solid phases present. At any rate, the appearance of a *mantle* about the top of a cumulus cloud (see figure 102), indicating the appearance of ice crystals, usually heralds the beginning of active rain. Cumulus clouds, even if well developed, which lack the ice crystal mantle, rarely cause precipitation. The mantle then, is a reliable guide as to whether rain and hail may be expected within cumulus clouds.

Lightning—The formation of lightning is apparently caused by the shattering of rain drops, due to violent agitation. A rain drop, which is initially neutral electrically, is probably separated into positively and negatively charged droplets when it is shattered. Generally the larger droplets have one electrical charge and the smaller droplets another. The larger droplets thus tend to fall out first, and leave an excess of electrostatic electricity of the opposite sign. This soon establishes a definite potential difference between various points within a single cloud, between individual clouds, and between clouds and the earth. When the potential difference becomes sufficient to break down the dielectric insulation of the air, or to cause progressive ionization of the air gap, a lightning flash results.

Lightning flashes are *direct current* discharges. The general impression is that they are alternating current discharges, but oscillograph records show conclusively that this is not the case. The amplitude of the voltage and current may vary considerably during a single flash but the current flow is essentially unidirectional.

No authenticated examples have appeared where all metal airplanes have been seriously damaged by lightning. Radio failures, involving the loss of antennae and the actual destruction of the radio apparatus have been reported. Damage to airplanes, however, has been limited to slight blackening of the metal where bonding was imperfect. Apparently, the airplane adjusts itself very rapidly to the potential of its surroundings by means of brush discharge. As a result, insufficient potential difference arises between the airplane and its surroundings to allow a sudden destructive discharge to strike it.

Even if a plane should be in the direct path of a lightning discharge it is probable that the current would pass through without damage, because of the highly effective bonding of modern airplanes.

TORNADOES

The tornado is a characteristic storm of the prairie region of the United States. It is the most violent and most sharply defined of all storms. The tornado may occur at almost any place in the central and southeastern portions of the United States. It rarely occurs north of latitude 40°. Tornadoes occur occasionally in other parts of the world within the general latitude limits of 20° to 40°, although they are very uncommon except in the United States. Occasionally, violent cold front thunderstorms are confused with tornadoes but the extremely high wind velocities of the true tornado are never experienced in ordinary thunderstorms.

Tornadoes generally first appear during the period of maximum convection during the mid-afternoon and they may continue actively well into the night. The horizontal wind velocities in an active tornado have never been measured accurately but, judging from the effects produced, they must amount to at least 300 to 400 miles per hour in many cases. The vertical velocities must also reach very high values since large objects are frequently carried a considerable distance in the air. Probably these vertical currents attain velocities of 200 miles per hour at times. The region of intense circulation associated with the actual tornado is generally not over a few thousand feet in diameter. Outside of this area of intense activity the winds may be strong but will never attain the tremendous velocity reached at the *eye* of the tornado. The storm itself generally moves with a velocity of from 20 to 40 miles per hour, and the length of the path traversed may be anything from a few miles up to several hundred miles. Often the base of the tornado cloud touches the ground and then rises into the air to descend again later, so that the path of the tornado on the ground is frequently discontinuous and irregular.

The synoptic situation associated with the occurrence of tornadoes is fairly well understood, although the exact localities which may be visited by the storms may only be forecast with a rough degree of accuracy.

First, a general condition of strong convective activity must be present, usually associated with the presence of Tropical Gulf or Tropical Atlantic air over the region.

Second, the presence of a slowly moving quasi-stationary cold front, generally indicated on the weather map as a sharp trough of low pressure, is also required. Whenever these conditions are met with in the great plains or southeastern portions of the United States, conditions are favorable for the occurrence of tornadoes.

The details of the atmospheric circulation which give rise to the tornado are very imperfectly understood. It is believed, however, that the tornado represents the extreme development of an unstable wave formed along a discontinuity surface (see page 226). The conditions for the development of such an unstable wave are provided along a quasi-stationary cold front in a region of pronounced convective activity. Apparently, under such conditions any wave which develops along the frontal surface tends to break down and become unstable. Under certain favorable distributions of airflow and pressure this unstable wave may rapidly increase in intensity until it reaches tornado proportions. Although the general synoptic situation required to give this type of storm is thus fairly well known the details from which reliable forecasts may be made are still imperfectly understood and apparently the same general situation may give rise to tornadoes at one time and fail to do so at another. The presence of the two synoptic features mentioned above, however, within the tornado regions of the United States should always be viewed with great suspicion by the forecaster.

Tornadoes, by virtue of their extremely high wind velocities and great turbulence are, of course, extremely dangerous for flying and all operations in a region affected by these storms should be suspended. During the day the typical tornado clouds may, of course, be observed and avoided, but at night there is no means by which the pilot can observe their presence. Flying under these conditions is extremely hazardous.

The energy required for the production and maintenance of a tornado is derived chiefly from the potential energy of mass distribution present between two masses of air, one considerably colder than the other. Thus is explained the invariable relation of the tornado to a cold front and its essential difference from the tropical hurricane which derives its principal source of energy from the condensation of large amounts of water.

HURRICANES

Occurrence—The tropical hurricane has been the subject of much study both as to the manner in which it occurs and the details of circulation which accompany it. Tropical hurricanes are true cyclonic storms which occur in certain rather limited portions of the tropics. They generally occur in the region between latitudes 10° and 30° in both the northern and southern hemispheres. They are not found in the immediate vicinity of the equator. They occur only rarely above latitude 30°, although occasionally they may reach 35° to 40° latitude, if they recurve. The tropical hurricane or *typhoon*, as it is called in the Far East, is confined to the open ocean and it invariably disintegrates rapidly as it passes inland. The area affected by the hurricane at any time may vary from 50 miles to nearly 1000 miles in diameter, and the cloud system of the storm may affect an area nearly 2000 miles in diameter. Generally the storm is approximately circular, and the rain area produced in connection with it is usually rather evenly distributed about the center. Exceptions occur when hurricanes move to the higher latitudes and draw relatively dry masses of air into their circulation.

Structure—The central portion, or *eye,* of the hurricane is marked by a region of comparatively clear sky with little rain and light winds. This eye may be from 5 to 25 miles in diameter and is marked by the region of lowest pressure. Extremely low pressures have been observed at the center of a hurricane, the lowest reading on record being 26.185 inches reported during a hurricane near the Philippine Islands in 1927. The pressure gradient near the center is often very intense and may give rise to extremely high wind velocities, which frequently attain 100 to 150 miles per hour. This region of high intensity is generally comparatively small. At distances of only 75 to 100 miles from the center of the disturbance the wind velocity is considerably less, and 150 to 200 miles from the center the velocities rarely exceed 40 to 45 miles per hour.

The velocity of movement of the center of the tropical hurricane is rather variable, ranging from a few miles per day in its earlier stages in the subequatorial regions, to approximately 100 miles a day after it has become well developed, and increasing to 300 or 400 miles per day after it has recurved. This *recurving* of the hurricane is very characteristic and is the normal result of the

storm's passing into the zone of prevailing westerlies after it has left the easterly winds of the trade wind belt. After the hurricane leaves the trade wind belt, and enters the zone of the westerlies it rapidly diminishes in intensity as it crosses cooler and cooler water.

Formation—Certain general conditions are necessary for the production of a tropical hurricane. These include a widespread area of nearly calm winds with active convection and general convergence. This region of light winds must be located several degrees away from the equator so that the earth's rotation may deflect inflowing air to produce active cyclonic circulation. Ascending currents in the convective region produce showers which release latent heat to provide for more convection. This in turn increases the rate of inflow of humid air at the surface, until a more or less steady state is reached with the deflective force balanced by the pressure gradient. Since the inflow proceeds spirally the conservation of angular momentum is an important factor in causing the very intense circulation near the center of the hurricane. The hurricane energy is supplied by the heat of condensation of ascending moist air that is derived from the immediate vicinity of the ocean surface. For this reason hurricanes rapidly dissipate on passing over land, or over cool waters.

The importance of convergent airflow in producing cyclogenesis has been clearly pointed out by Petterssen (see page 217). Convergence along the so-called *equatorial line of convergence* in the southwestern north Pacific is probably the principal cause for the formation of typhoons in the Far East. H.-C. Huang has shown recently that typhoons can be traced very definitely to this line. When the Pacific HIGH moves far to the east of its normal position, a zone of light variable winds appears near latitude 15°, just north of the doldrums and east of the Philippine Islands (figure 41). This region corresponds closely with the *equatorial line of convergence,* where Equatorial air from the southeast meets Tropical air from the east and east-southeast. This "front" was recognized by Werenskiold and Bjerknes some time ago. Actually it is not a *front,* since the solenoidal field is insufficiently concentrated to warrant that designation. It is very definitely a zone of convergence, however, as Huang shows.

As the Pacific HIGH commences its return westward, the convergence along *equatorial zone of convergence* increases and cyclonic

circulation rapidly develops (page 217). Since abundant energy is available in the very warm and moist equatorial and tropical air, convection rapidly increases. This leads to still further cyclogenesis and still greater inflow of convectively unstable air. A very intense cyclone center rapidly develops and the fully formed typhoon soon appears.

The conditions leading to the formation of the tropical hurricanes of the southwestern north Atlantic are practically identical to those described for the north Pacific. A general zone of convergence exists there between Equatorial air from the southeast and Tropical air from the east-southeast. The Tropical air is usually of relatively recent Polar origin, while the Equatorial air has been over the equator for many days. This situation is also similar to that in the Far East. As the Bermuda-Azores HIGH moves eastward from its normal position a region of relatively low pressure is developed in the eastern Caribbean Sea area. When the HIGH returns westward convergence increases along the Tropical-Equatorial boundary and cyclogenesis takes place very readily with the rapid formation of a hurricane. This proceeds westward, then northwestward, and finally recurves and passes into the zone of prevailing westerlies becoming an extropical cyclone.

The general pressure situation leading to the formation of hurricanes in both the Pacific and Atlantic regions is fairly well defined. The areas affected by hurricanes are all on the west side of the great subtropical high pressure belts. They thus correspond with the region of maximum vertical amplitude of particles moving around the HIGHS (see page 88). *Mechanical lifting* of particles moving from east to west in these regions is thus apparently a potent force in releasing the convectional instability in the Equatorial and Tropical air. The necessary combination of *convergence* along the "equatorial lines of convergence," and *lifting* as particles move from east to west, is attained in only a few regions on the earth's surface. These regions are the "breeding grounds" of practically all hurricanes.

Hurricane Movements—In the latitudes under discussion true *fronts* do not exist, since the temperature contrasts necessary to produce a concentration of the solenoidal field are lacking. The typhoon is thus a pure vortex motion in essentially homogeneous air. Its motion, in the early stages is determined by the *convective component* of the general circulation, since the dynamic component is

practically negligible near the equator. Thus, it moves westward during its earlier stages. As the storm moves farther north (or south, in the southern hemisphere) the dynamic component increases until it balances the convective component. At this point, the storm moves directly away from the equator for a short time until the dynamic component exceeds the convective, and the storm recurves toward the east.

After the storm has recurved and passed out of the tropics it may become related to extratropical disturbances. Fronts may then appear and air masses of varying properties may enter the typhoon's circulation. The storm then becomes an ordinary extratropical cyclone and behaves in all respects like one.

Within the trade wind belt hurricanes generally move with a comparatively uniform velocity and follow a fairly regular path so that their course can be plotted with considerable confidence, once it is established. After they move to higher altitudes, however, they are generally subject to the conditions which affect extratropical cyclones and their path becomes more difficult to plot. They invariably tend to recurve to the north and northwest (in the northern hemisphere), but the point at which this recurving process begins depends upon the general pressure field existing at the time, as well as upon the air masses which are present in the region, and is thus rather difficult to determine.

The early view, that a hurricane affects only the surface stratum, has been shown to be wholly erroneous by a careful study of its structure. Actually it affects the weather conditions at a great distance above the surface, probably involving in most instances the entire region below the tropopause. For this reason it is impracticable to fly over such a disturbance and it must therefore be avoided by pilots under all circumstances. Since the velocity of movement of a hurricane is comparatively slight there is generally no reason for the pilot not having ample warning of its presence.

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CHAPTER 13

CONDENSATION AND PRECIPITATION

INTRODUCTION

It is important in discussing clouds and rain or snow to distinguish carefully between the meteorological processes of *condensation* and *precipitation*. Condensation may be defined as the process by which invisible water vapor in the atmosphere is changed to a visible cloud of water droplets. Precipitation on the other hand is the falling out or removal of water or ice particles from such a visible cloud. Each of these phenomena is a distinct physical process, and each is the result of different causes. In other words, the physical causes of condensation are not able by themselves to cause active precipitation; and the processes which cause active precipitation are not effective alone, but must have condensation products on which to act. Recent investigators, including A. Schmauss, A. Wigand, and T. Bergeron, have shown that clouds may be considered as *colloids*. The water droplets or ice crystals of a cloud thus are suspended in the surrounding air and obey the laws governing colloidal suspensions. This conception of the structure of clouds has been very important in recent theories concerning precipitation and its causes.

CONDENSATION NUCLEI

It has been well established that the condensing of water vapor in the atmosphere into definite, though minute, water droplets, demands the presence of a *nucleus* for each droplet so formed. These nuclei in atmospheric condensation are minute foreign particles of a size ranging from 10^{-8} to 7 microns in radius. Most of them are thus too small to be studied directly, even with the most powerful conventional microscope, since all but the largest are smaller than the wave length of even ultra-violet light. Indirect evidence, however, shows clearly that they are present in the atmosphere, and gives us a fair estimate of their sizes. It is to be hoped

that investigation with the newly developed electron microscope, which is capable of resolving particles as fine as 5×10^{-8} micron in diameter, will reveal definitely the size and character of nuclei. With perfectly pure, washed air, in which no foreign particles are present to act as nuclei, it has been found that a supersaturation of at least 420% is necessary to cause condensation. At this value of supersaturation apparently the negative ions in the air begin to act as condensation nuclei. At the even greater supersaturation of 600% the positive ions begin to act as condensation nuclei, and at 790% all ions are active. Since in the atmosphere, the degree of supersaturation never exceeds about 110%, it is clear that foreign particles must act as nuclei of condensation in all cases in nature.

It has been found that certain foreign substances are much more effective nuclei than others. These are the hygroscopic, or water-absorbing substances. Sulphuric acid is especially important in nature, since it is not only one of the most hygroscopic of materials, but also because it is very abundant wherever combustion is occurring. Sulphuric acid is produced by the oxidation of SO_2, a very common compound in polluted air. This oxidation usually requires the influence of ultra-violet light, and takes place through the following chemical reactions:

$$3SO_2 + O_3 \rightarrow 3SO_3$$

$$SO_3 + H_2O \rightarrow H_2SO_4$$

or

$$SO_2 + H_2O_2 \rightarrow H_2SO_4$$

J. Aitken first pointed out the importance of sunlight in producing active condensation nuclei by means of the above reactions. Apparently the ultra-violet radiation in sunlight oxidizes the oxygen in the atmosphere to produce either ozone or hydrogen peroxide. These strong oxidizers then act on the SO_2 in the air. Many other hygroscopic substances have also been shown to act as nuclei. Thus, sodium chloride and other salts from sea spray are important over and near the ocean. Compounds of nitrogen, formed through the action of ozone, are also of some importance. Many other hygroscopic materials are of probably minor importance as nuclei. Dust particles are probably only rarely of importance as condensation nuclei, since other more active materials usually furnish all the nuclei needed in the atmosphere.

The number of condensation nuclei in the atmosphere varies tremendously. H. Landsberg points out that there are a definite diurnal and an annual variation in nuclei number, and that humidity, wind, type of air mass, electrical potential, radio-activity, and other meteorological elements all tend to influence the number of nuclei present. Very pure air in nature may exhibit as few as a few hundred condensation nuclei per cc, while heavily polluted air may show several million nuclei per cc. These measurements, which were all made with some form of nucleus counter, probably included a fairly high percentage of large ions acting as nuclei at rather high supersaturations. Thus, in very pure air, such as might be found in mountainous regions there may be actually only a very few active nuclei available. In such cases there may thus be a definite deficit of nuclei. In general, however, there are usually ample nuclei present for all condensation processes. The common range in the number of nuclei which will become active at slight supersaturations only is perhaps 10 to 10,000 per cc.

SUBLIMATION NUCLEI

The process of *sublimation*, in which water vapor passes directly over to the ice stage without the intervention of the liquid stage, is thought by T. Bergeron and W. Findeisen to be of considerable importance in the atmosphere. They think that the sublimation process requires nuclei just as the condensation process does, although the sublimation nuclei probably are of a different type from the hygroscopic nuclei of condensation. They reason that the hygroscopic condensation nuclei are relatively strong salt solutions while they are of nuclear sizes, and therefore probably remain liquid far below 0° C. (due to the molecular depression of the freezing point). As a result of this, they cannot serve as nuclei for the formation of ice crystals. It is thought rather that minute mineral grains, particularly those of the same crystal system as ice (hexagonal) such as quartz, serve as nuclei of sublimation. These nuclei are probably of about the same size as the condensation nuclei (10^{-3} to .7 microns in radius).

Sublimation nuclei are apparently much less abundant in the atmosphere than condensation nuclei. In fact it is believed that the number of sublimation nuclei amounts to from 1/100 to perhaps 10 per cc. From this it may be seen that there may be times when

not enough sublimation nuclei are available to produce ice crystals. This may be of considerable importance during the rejuvenation on old clouds as will be seen later.

CONDENSATION IN THE ATMOSPHERE

The process of condensation in the atmosphere is a continuing one. It occurs when the relative humidity rises (due to lifting, radiation, mixing) until saturation is nearly reached. When extremely hygroscopic nuclei, such as those composed of sulphuric acid, are available, condensation may commence at comparatively low relative humidities. The nuclei then grow slowly, becoming dilute solutions as the humidity increases. This phase of the condensation process results in the formation of haze. During this stage, the growth of droplets is very slow, as may be seen from figure 88. From this figure it may be seen that an increase in vapor pressure from 3.6 mm. (78% R.H.) to about 4.4 mm. (96% R.H.) will result in but a slight increase in the radius of a sodium chloride nucleus having a mass of 10^{-17} gram, whereas any increase above 4.4 mm. results in a very rapid increase. The curves in this figure also show that as the saturation point is approached, condensation will take place first on the larger nuclei. The smaller nuclei (such as the sodium chloride nuclei with a weight of 10^{-19} gram) require an appreciable degree of supersaturation and presumably will not become active until the larger nuclei have all been utilized. Another important fact that may be seen from figure 88 is that although condensation starts on the larger hygroscopic nuclei at fairly low humidities, these nuclei do not grow to cloud particle size (about 1 micron in radius or larger) until the humidity reaches a value sufficient to cause condensation on neutral nuclei of the same size. However, so many hygroscopic nuclei have become active by this time that it may be presumed that a large percentage of all cloud droplets have hygroscopic particles for their nuclei.

In spite of the selective tendency mentioned in the preceding paragraph, condensation will usually proceed on nuclei of widely differing sizes, since this effect is rather feeble. However, there is another effect which tends to cause droplets of the same *age* to become uniform in size as condensation proceeds. H. G. Houghton points out that if two growing drops are initially 0.2 and 2.0 microns in

diameter, the larger will be only 10.2 when the smaller has grown to 10.0 microns. He obtained this result from the expression:

$$a^2 = A^2 + 8k\ (D - D_0)\ t$$

where a is drop diameter after time t, A is initial drop diameter, k is diffusion coefficient of water vapor in air, and $D - D_0$ is difference between water vapor density in the atmosphere and at the drop

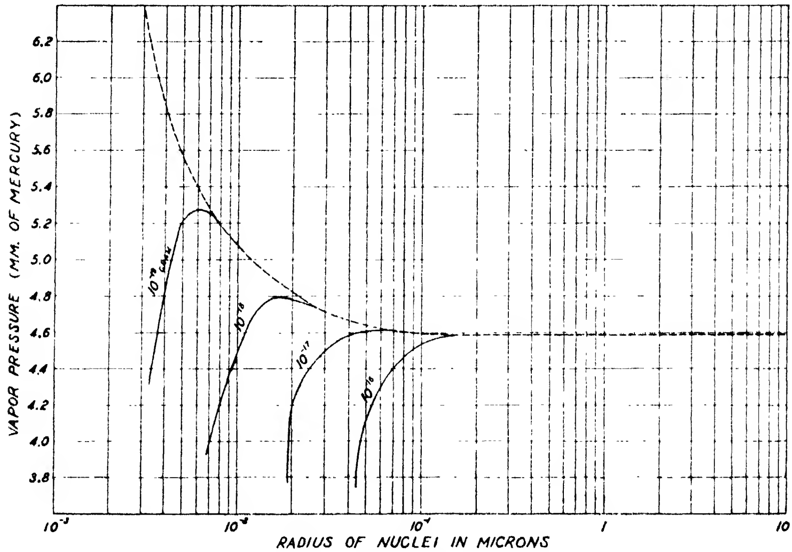
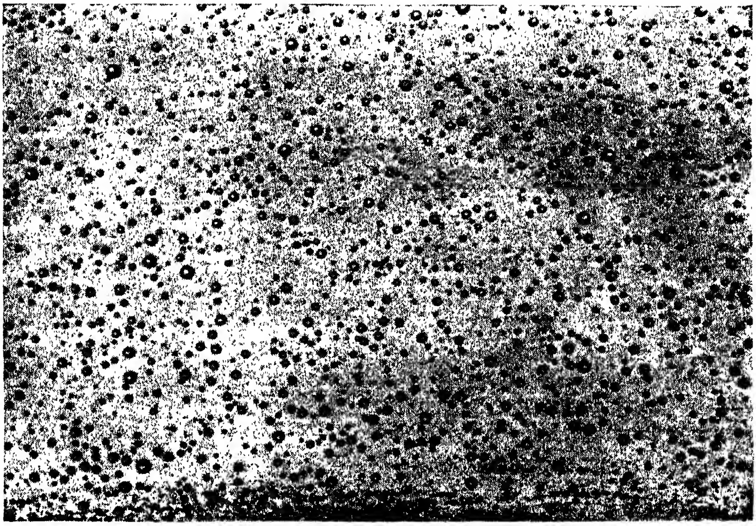


FIGURE 88.—CURVES SHOWING THE SUPERSATURATION REQUIRED TO INDUCE CONDENSATION ON SPHERICAL NUCLEI AS A FUNCTION OF THEIR RADII

The dashed curve is for neutral or dust nuclei, while the solid curves apply to nuclei of sodium chloride of dry weights as indicated. Computed for a temperature of 0° C. and a saturation vapor pressure of 4.59 mm. of mercury. (After H. Kohler and H. G Houghton.)

surface. Although this is true, nevertheless differences in the *age* of droplets may lead to wide size differences within a comparatively small region through the operation of other processes, as will be seen later.

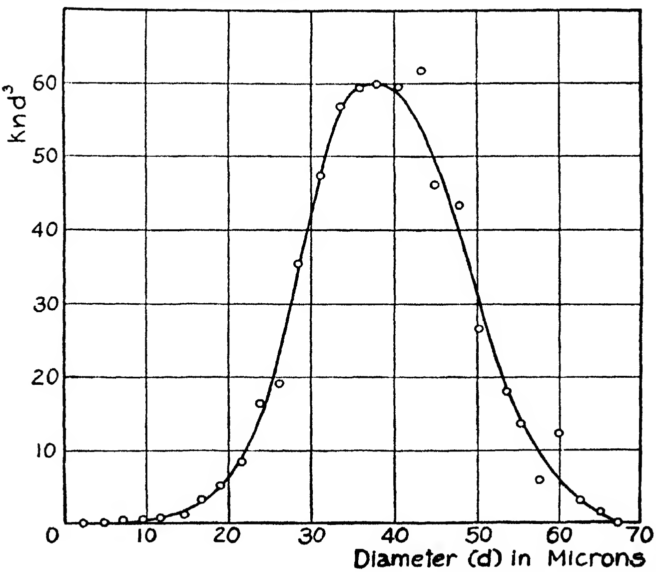
Condensation processes alone are able to produce only relatively small droplets—perhaps 50–200 microns in diameter. Droplets of this maximum size correspond with large cloud particles (see figures 89 and 90) which range up to perhaps 100 microns, and small drizzle particles. [Drizzle is defined by the U. S. Weather Bureau in Circular N, as consisting of minute droplets having a diameter less than



Courtesy H. G. Houghton

FIGURE 89.—PHOTOMICROGRAPH OF NATURAL FOG PARTICLES

Made at Round Hill, Mass., Feb. 17, 1936. Magnification of drops as here reproduced (including flattening effect) about 32 times.



Courtesy H. G. Houghton

FIGURE 90.—VOLUME DISTRIBUTION CURVE FOR ABOVE FOG SAMPLE

The quantity knd^3 used as one of the coordinates in this figure is a function of the number of drops n corresponding to particles of a diameter d, and an arbitrary constant k. Thus, the curve shows the distribution of liquid water among the particle sizes present in the fog. 1621 drops were measured to obtain data for this curve. The peak of the curve is near 38 microns.

1/50 inch (508 microns.)] Drops larger than about 200 microns thus must be the result of the action of other processes than pure condensation. These will be discussed under Colloidal Stability in Clouds.

WATER CLOUDS

After saturation is reached in the atmosphere, the relative humidity continues to increase, and the nuclei continue to grow slowly until a certain critical humidity is attained. Normally this occurs at relative humidities between 101 and 106%. After this value is reached, the nuclei can increase in size without any further increase in the relative humidity.

It is believed that no new droplets will be formed after this critical humidity is reached, and that only those already formed will continue to grow. If this is true there is an important relationship between the rate of cloud formation and the number of drops formed. With rapid condensation, the critical relative humidity is reached, and may even be materially exceeded for a time before equilibrium can be established. As a result, very many nuclei become active, and large numbers of droplets are formed. With slow condensation, however, the relative humidity attains its critical value after a slow rise so that comparatively few, but relatively large, droplets are formed.

As a result, in cumulus activity, even with only moderate vertical velocities (3 to 5 meters per second), the relative humidity may reach 102% to 112%, and as many as 10,000 drops per cc may be formed. In layer clouds on the other hand, the relative humidity rarely exceeds 100% and the number of drops per cc usually amounts to about 100 or so.

Figure 89 shows clearly the rather wide range in droplet size that may be found in an ordinary stratus cloud (fog). The heterogeneous size distribution indicated by this figure may probably be taken as being more or less typical of conditions in stratiform clouds. In cumulus clouds the sizes are probably even more heterogeneous because of the turbulent mixing constantly taking place. These observations are actually not contradictory to the fact mentioned earlier, that droplets of the same age tend toward uniformity of size. Actually it merely indicates that droplets of widely differing *ages* are found side by side in a cloud. There is also a very slight tendency

for the larger droplets in a cloud to grow at the expense of the smaller, due to the higher vapor pressure over the smaller droplets resulting from their smaller radii of curvature. This may still further tend to increase the heterogeneity of the droplet distribution.

Since, as has been pointed out above, condensation alone can produce droplets having diameters of only 50–200 microns, further increase in size can proceed only through coalescence. The fact has already been mentioned that clouds may be considered to behave as colloids. It is believed that the air acts as the dispersing medium, and that the water droplets (or ice particles in the case of ice clouds) act as the disperse phase. If this is true, then the laws governing colloidal stability may be applied to clouds.

COLLOIDAL STABILITY IN CLOUDS

Colloidal stability may be regarded for the case of cloud particles simply as "the lack of tendency to coalesce." When colloidal stability disappears, or the clouds become colloidally unstable, active precipitation may ensue due to coalescence. Bergeron points out that colloidal stability in clouds is promoted by the following factors:

1. Uniform (equal, unipolar) electric charge, causing repulsion, and thus counteracting the fusion of droplets.

<table>
<tr><td>2. Uniform size of cloud elements.

3. Uniform temperature of cloud elements.</td><td>Either or both causing equal surface tension and thus equal vapor pressure over droplets. This prevents diffusion transport of water from smaller (or warmer) droplets to larger (or colder).</td></tr>
</table>

4. Uniform motion of cloud elements, preventing fusion by collision.

5. Not over two phases of water (solid, liquid, or vapor) present.

In newly formed clouds, (1) and (2) may be expected to hold from reasons of symmetry; (3) should prevail at least in the interior of such clouds; (4) should hold in the free atmosphere when turbulence is lacking. Condition (5) will be discussed later.

On investigation it is found that the first four conditions neces-

sary for colloidal stability outlined above occur in nature in most cases. The effect of non-uniform electric charge in promoting droplet coalescence may be of some importance in thunderstorms *after* droplets have already grown to precipitation element size, but as a primary means of destroying colloidal stability it is of no practical importance.

The effect of non-uniform size in promoting colloidal instability is to cause a vapor pressure differential from the smaller to the larger droplets. For droplets less than about 2 microns in diameter (the size of the smallest elements of dry fog) it is appreciable. For ordinary rain drops or snow flakes it acts perhaps a billion times too slowly.

The effect of non-uniform temperature causes a vapor pressure differential to be directed from the warmer droplets to the cooler ones. These differences in temperature may arise as a result of insolational and radiational effects or because of different origins of the droplets. As early as 1877, O. Reynolds suggested that uneven illumination of cumuliform clouds could cause precipitation. He reasoned that a strong vapor pressure differential would be established, so that relatively cool droplets in the shade would grow actively at the expense of nearby droplets in the direct sunlight, and thus reach precipitation size. Certain types of precipitation probably do owe their occurrence of Reynolds' theory, such as occasional showers from stratocumulus clouds in the late afternoon, and from cumulus (not cumulonimbus) clouds in the tropics late in the day. The difficulty with this theory as a *general explanation* of precipitation is that rain or snow often falls from both stratiform and cumuliform clouds at times when uneven illumination does not exist.

Temperature differences between adjoining droplets could also be caused by their originating in regions of widely differing temperatures. In this way the cold drops could grow at the expense of the warmer ones in their vicinity. Actually however, appreciable temperature differences between adjoining droplets are rarely brought about in this manner. Only in cases where droplets have different velocities of fall can this process be of real importance, but even here the *initial* size difference must be brought about in some other manner. Thus, temperature contrasts may *contribute* to precipitation, but cannot *initiate* it.

Turbulence may be of some importance in connection with (4), at least near the ground where it may aid in releasing drizzle. In

the upper levels of the atmosphere, above about 600 meters, however, the air flow approaches a laminar condition, at least on a small scale, so that collisions between droplets are not very effective in causing coalescence. It should be pointed out here that neither Bergeron or Findeisen attach very much weight to a process which other writers think is of considerable importance in the growth of cloud elements. This is the fusion of drops, as larger ones fall by gravity and coalesce with smaller ones by impact. H. G. Houghton thinks this process is of major importance in the growth of cloud droplets and rain drops. Findeisen thinks that hydrodynamic effects between droplets of widely differing sizes effectively prevent their coalescence, although Houghton mentions experiments in which such an effect is not apparent.

Bergeron sums up the practical effects of (1), (2), (3) and (4) by saying, ". . . none of the hitherto recognized factors of cloud coagulation represents the *universal* release of *real* precipitation (real rain or snow). Either the factor is only active at special times of day (3), or under abnormal electric conditions (1), or can only coagulate droplets of fog dimensions or smaller, without causing few large drops (1), (2), (4), or is ineffective on the whole in the *release* of coagulation."

Bergeron then turns to a discussion of condition (5), namely the effect produced when both liquid and solid phases are present with water vapor. He shows that this factor can have an effect of real magnitude, and is capable, he thinks, of initiating and promoting the rapid growth of certain cloud elements (droplets or crystals) at the expense of others. This effect is the result of the different vapor pressures which exist over ice and subcooled water at the same temperatures. For example, at $-10°$ C., the vapor pressure over water is 2.859 mb., while over ice it is 2.619. Thus at this temperature, the vapor pressure over water is 0.240 mb. higher than over ice (see figure 91), and a strong tendency therefore exists for water to be transported from a water surface to an adjacent ice surface. Because of this tendency a mixture of the three phases—ice crystals, liquid water drops, and water vapor—constantly tends to be transformed into the two-phase system of ice crystals and water vapor.

The temperature to which water may be cooled in the atmosphere without freezing varies with the amount of agitation of the droplets. It is Bergeron's belief that the crystallizing forces acting on subcooled water droplets increase with decreasing temperatures,

so that greatly subcooled droplets require less agitation to initiate crystal formation than slightly cooled ones. This matter of the various factors involved in the freezing of subcooled water has been rather imperfectly studied experimentally. For instance, very little is known about the effect of time, yet this very probably is of considerable importance in the water-ice equilibrium. Findeisen thinks

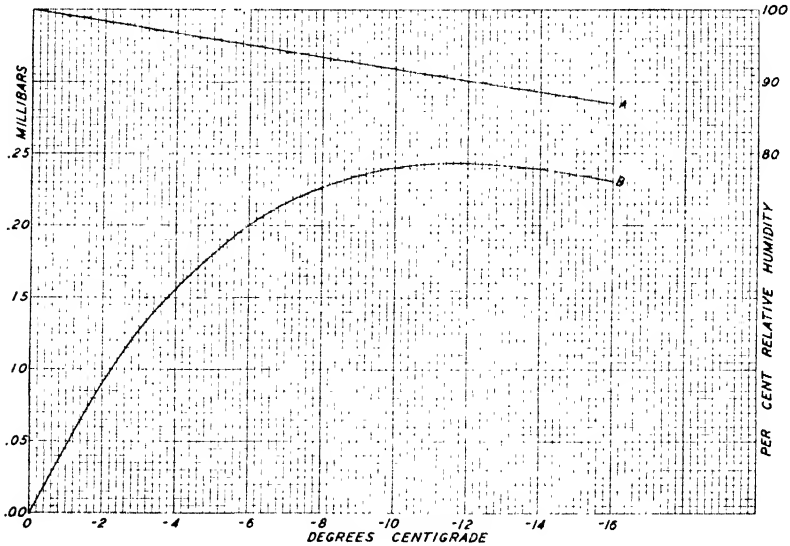


FIGURE 91

A—Curve showing relative humidity over an ice surface at various temperatures compared to a subcooled water surface over which the relative humidity remains at 100%.
B—Curve showing vapor pressure differential between subcooled water and ice at various temperatures. (Data for both *A* and *B* obtained from Smithsonian Physical Tables, 8th Ed.)

that all ice crystals arise from direct sublimation on sublimation nuclei, and never by direct transformation from the liquid to the solid state. This may be too extreme a view, for it is quite possible that water droplets may change spontaneously to ice crystals if given sufficient time, even in the absence of sublimation nuclei.

ICE CLOUDS AND WATER-ICE CLOUDS

After ice crystals have once been formed on sublimation nuclei, they continue to grow by direct sublimation from the water vapor present. If liquid water is present, the sublimation process will proceed very rapidly, because in this case a strong supersaturation

exists over the ice surface. At $-10°$ C. the vapor pressure difference between water and ice amounts to 0.24 mb. as was pointed out above (see figure 91). This is equivalent to a supersaturation over ice of 9%; at $-16°$ C. the supersaturation amounts to 13% (see figure 91). Actually some supersaturation is necessary before ice crystal formation commences on the sublimation nuclei—perhaps 10–15%—since they do not become active at comparatively low humidities as do some condensation nuclei. Beyond this value of the humidity, however, the tendency of the nuclei to become active is very strong, and once crystallization has started it proceeds readily. As long as any liquid water is present, supersaturation is maintained over the ice crystal surfaces, and rapid growth continues.

The forms which ice crystals take as they grow depend largely on the degree of supersaturation which exists. With slight supersaturation, simple ice needles or plates are formed. These appear in a variety of forms all exhibiting the characteristic hexagonal crystallization of ice. Of frequent occurrence are elongated hexagonal prisms, often terminated at one or both ends with hexagonal pyramids. These average perhaps 1 mm. in length. Also very common are small tabular hexagonal plates. These ice needles and plates comprise such ice clouds as cirrus and true altostratus. Halo phenomena in these clouds are caused by the regular arrangement of the ice crystals or plates composing them. As a result of this even orientation of the cloud elements light is refracted uniformly in passing through them. These crystals and needles are too small to act as precipitation elements except in their larger sizes.

As the degree of supersaturation increases, the corners of these simple ice crystals shoot out to form elementary six-pointed stars, the skeletons of snow flakes. These may develop into snow flakes of almost unending diversity of detail, all however showing the basic hexagonal form. With further supersaturation the crystal branching increases in intensity, particularly in directions at right angles to the plane of the simple snow flakes, yielding complex, more or less spherical aggregates of snow flakes and ice needles. These are commonly called *snow pellets* or *graupel.* They are of very delicate structure, and usually break into several pieces on striking the ground. Coalescence between ice needles and snow flakes also contributes to the complexity of these forms.

With still greater supersaturation, subcooled water droplets are usually present. These impinge on the falling snow pellets, freeze

because of the presence of an ice nucleus, and cement the corners together to form the well-known *sleet* particles. Sleet grains are fairly compact, and usually rebound from hard surfaces without breaking. They have formerly been assumed to result from the freezing of rain drops as they pass through a cold layer of air near the surface. If Bergeron's and Findeisen's views are to be accepted however, the spontaneous freezing of rain drops is impossible except on impact with a solid object, so that this older idea seems now quite untenable. Even if other features of the Bergeron-Findeisen hypothesis cannot be fully accepted, the presence of subcooled water in the atmosphere has been so generally accepted, that the formation of sleet through the freezing of rain drops as they pass through cold air seems definitely disproved.

As the supersaturation increases even further, and more and more subcooled water droplets are present, *hail* may form. Hail and sleet, both of which require very high supersaturation, only arise under conditions of very rapid adiabatic cooling, usually a result of strong convection. In such cases, because of the high vertical velocities developed, the relative humidity in a cloud may greatly exceed 100%, due simply to the fact that the water drops or ice crystals cannot grow as rapidly as water vapor is made available.

It was mentioned earlier that the number of sublimation nuclei is usually much less than that of condensation nuclei (perhaps 1/100 to 1/1000 as many). Largely because of this, the number of cloud elements in ice clouds is usually much less than in water clouds. Visibility in ice clouds is consequently considerably better than in water clouds. The particles in ice clouds generally are much larger than those in water clouds, because there are fewer of them to receive the available water vapor.

The rate of fall of ice crystals is usually rather high, due to their fairly large size. They may thus fall a considerable distance into the dry air below them before they evaporate, whereas the small water droplets in water clouds tend to evaporate rapidly. For this reason water clouds exhibit clean edges, whereas ice clouds usually have ragged edges. Even cirrus clouds show by their wispy nature that relatively large particles exist at very high levels. This is especially interesting, since it shows that very few sublimation nuclei can be present at cirrus levels. Because of this the extremely small amount of water vapor which is available can form large crystals which, in turn, cause ragged cloud edges.

Occasionally, water drops may condense in or near ice clouds. This may occur when sublimation has proceeded for so long a time that all of the sublimation nuclei are used up. Continual increase of the relative humidity then leads to condensation on the condensation nuclei, even though the saturation point may not be reached until the temperature falls far below 0° C. Such a process often occurs just above old ice clouds (altostratus or nimbostratus) in regions where the sublimation nuclei have all been used up. This usually happens when convection is resumed after a period of quiescence, and results in the formation of an altocumulus cap on top of an old altostratus or nimbostratus cloud.

PRECIPITATION

The size above which cloud elements become precipitation elements may be placed conveniently at a diameter of about 10^{-2} cm. Below this size, droplets may fall out of clouds, but they usually evaporate before reaching the earth. Of course the size which falling droplets must have to reach the earth depends on the cloud height and on the relative humidity of the air between the cloud and the earth. It might be mentioned that the distance which a droplet can fall through unsaturated air without completely evaporating is directly proportional to the *fourth* power of the radius. As a consequence of this, a slight increase in a droplet's size allows it to fall much farther before evaporating.

In pure water clouds, droplets can grow either by condensation or through coalescence. Normally, neither of these processes can give rise to large drops. Even with very few nuclei, and with very rapid cooling, droplets of only a size less than 10^{-2} cm. can be produced by condensation alone. The importance of coalescence has been widely debated as mentioned earlier. Findeisen thinks it is of very minor importance, whereas Houghton believes it to be of considerable importance. It will suffice here to point out that large drops are only very rarely observed in connection with pure water clouds. Findeisen's conclusion therefore appears to be the more tenable in the light of this and other evidence. Precipitation from water clouds occurs only as drizzle from low lying clouds. With water clouds at considerable heights, precipitation is only rarely observed, since the falling droplets usually evaporate before reaching

the earth. Exceptions should be noted in which showers may occur from cumulus clouds in the late afternoon due entirely to temperature contrasts caused by uneven illumination. This process may also cause light precipitation from the cloud type *stratocumulus vesperalis*, in the late afternoon or early evening.

With pure ice clouds, relatively few large ice crystals tend form in the sublimation process, due to the fact that usually only a few sublimation nuclei are available. If sublimation is maintained by continuous adiabatic cooling, these ice crystals may grow to a rather large size, passing through the snow crystal stage and occasionally reaching the snow pellet stage. These elements may be considerably larger than the 10^{-2} cm. limit between cloud and precipitation elements, so that active precipitation may occur. The probability of active precipitation is thus considerably greater with pure ice clouds, than with pure water clouds. Coalescence due to impact caused by varying rates of fall of the ice crystals may also aid in the formation of snow flakes and snow pellets. This process is most active at temperatures near 0° C., since at lower temperatures there is little tendency for the separate elements to fuse. Rain may of course result from the melting of snow flakes which fall from an ice cloud. With a large number of particles, the tendency for them to coalesce and form large elements is considerable. With a small number of particles it is slight. Therefore, the size of the elements which fall from an ice cloud is directly proportional to the intensity of the precipitation.

It should be emphasized that ice clouds occur in the atmosphere much more frequently than has been thought previously. Actually they are probably encountered more frequently than water clouds in the temperate zone. Typical altostratus is the usual precipitation producing ice cloud. This cloud, in fact, is in every way equivalent to the nimbostratus form, except in vertical extent. Both of these cloud types occur in connection with nearly every low pressure system, and they are the major cause of most widespread precipitation. The importance of ice clouds for precipitation, either acting alone, or together with water clouds, cannot be overemphasized, according to Findeisen.

Markedly high precipitation intensities, however, do not occur in connection with pure ice clouds. Rather, heavy rain and snow usually fall from clouds in which ice crystals occur side by side with subcooled water droplets. In these complex clouds, the ice crystals

tend to grow very rapidly, as has been pointed out, at the expense of the water droplets. In such a manner very large ice crystal aggregates may be formed. This growth of the ice crystal forms takes place much more rapidly in complex clouds than in pure ice clouds, for here adiabatic cooling, which is alone effective in pure ice clouds, is greatly aided by the different vapor pressures over the two phases, ice and water. In falling through the cloud, these aggregates—in the form of snow pellets, sleet grains, or hail stones—continue to grow by coalescence with ice crystals and water droplets which they encounter.

PRECIPITATION FROM STRATIFORM CLOUDS

Figure 92 represents a generalized cross-section of conditions which may occur in stratified clouds, with both ice crystal and water droplet clouds present. The main cloud mass is a nimbostratus type composed entirely of ice crystals. To the right are some stratocumuli, composed of subcooled water droplets, and below the 0° C. isotherm are a few ragged fractostratus or scud. Moderate rain occurs only in regions where ice crystals fall into water clouds and thus cause the rapid growth of precipitation elements, due to the presence of solid and liquid phases side by side. Light rain occurs where falling ice crystals melt. A light drizzle may occur beneath the water clouds which have not been "infected" by ice crystals, where such clouds are low enough. The altocumuli (water clouds) at the top of the nimbostratus result from renewed convection in the upper portion of the ice cloud, in a region where insufficient sublimation nuclei are present to yield ice crystals. These clouds are strictly secondary to the main nimbostratus mass, although they occur rather frequently. They are usually 200-300 meters thick, although they may attain considerably greater thicknesses at times.

PRECIPITATION FROM CUMULONIMBUS CLOUDS

Figure 93 represents the life history of a cumulonimbus cloud. This originates as a pure water cloud (cumulus humilis, becoming c. congestus), and continues as such considerably above the 0° C. isotherm. As it continues its upward growth however, the upper portions finally reach a critical temperature at which sublimation commences directly on the sublimation nuclei. (Until this is reached the

supersaturation of the water vapor with respect to ice is not great enough to initiate sublimation.) This generally occurs at a temperature between $-15°$ and $-20°$ C., or perhaps 3000 meters above the 0° C. isotherm (assuming continuous moist adiabatic reduction of

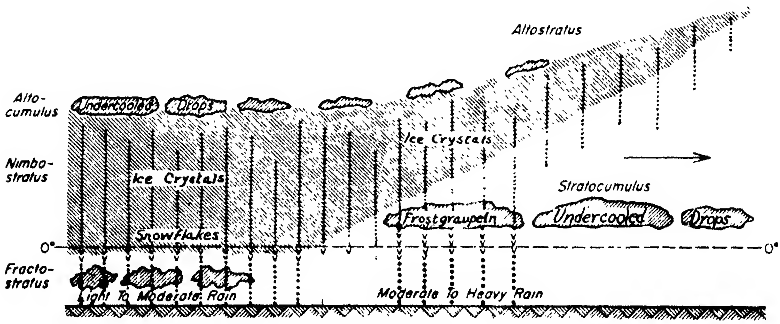


FIGURE 92.—PRECIPITATION FROM NIMBOSTRATUS CLOUD SYSTEM

Showing distribution of ice and water clouds and areas of light, moderate, and heavy rain in a nimbostratus cloud system. The term "frostgraupeln" is apparently equivalent to the U. S. "sleet particles." (After Findeisen.)

temperature). Above this critical level ice crytals form very rapidly and the undercooled water drops vanish with equal rapidity by evaporation. Snow pellets and sleet are formed.

When these attain a size such that they can no longer be supported by the vertical currents which are present, they fall out of

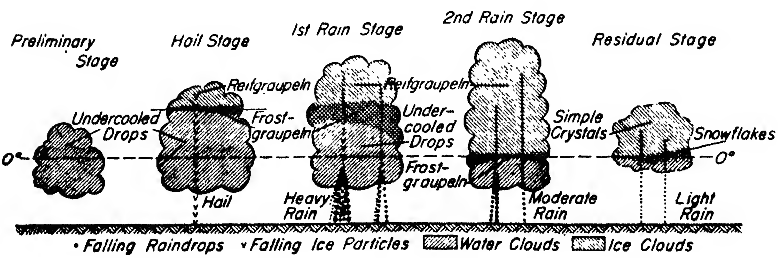


FIGURE 93.—PRECIPITATION FROM CUMULONIMBUS CLOUDS

Showing growth and decay of cumulonimbus cloud with various types of precipitation After Findeisen.)

the ice crystal region and penetrate the undercooled water clouds lying below the critical level. There they continue to grow by impact, as well as through the vapor pressure differential between the two phases, ice and water. They may now grow into large hail

stones, which may fall rapidly enough to reach the ground without melting. The size of the hail stones depends on the distance between the 0° C. isotherm and the critical temperature, and on the velocity of the vertical currents tending to support them.

The hail stage of the cloud is succeeded by the first or heavy rain stage. This ensues when the portion of the cloud composed of subcooled water droplets diminishes as more and more of the droplets are adsorbed by falling ice aggregates. Finally insufficient water droplets remain to yield large hail stones, and the snow pellets and sleet grains themselves fall out of the cloud. As these melt below the 0° C. isotherm, they usually produce several large rain drops. This production of several drops from one solid particle results from the fact that the size of a falling water drop cannot be greater than about 5 mm. Above this size it breaks up into smaller drops. The falling snow pellets and sleet grains, however, may have a much greater size than this since they are solid particles. It is believed that this atomizing effect probably has an important effect in producing thunderstorm electricity. (See page 248.)

As precipitation continues, practically all of the cloud elements above the 0° C. isotherm become ice crystals, and in the lack of subcooled water droplets, the formation of large precipitation elements becomes impossible. Raindrops are now formed only by melting ice crystals, which are now much smaller than at first, and grow mainly by impact with the warm water droplets below the 0° C. isotherm (second rain stage).

Finally, the water portion of the original cumulonimbus cloud practically disappears, and only a light secondary rain persists (residual stage).

The length of time required for the complete life history described here varies greatly, of course. It depends primarily on the intensity and persistence of the vertical currents which produce the cumulonimbus cloud. With strong tropical storms, the vertical currents persist for many hours, and the hail stage may not ensue for several hours after the beginning of convection on account of the great height of the 0° C. level. After the hail stage commences, however, it may last for a long time, and the storm may persist all day. With small thunderstorms, on the other hand, the hail stage may not occur at all, and the rain stages may be passed over quickly. In small shower clouds the ice crystals, in the absence of strong vertical currents, remain so small that when they melt, little atom-

izing of the drops ensues. As a result, electrical phenomena in these clouds are only feeble.

Sometimes interruptions in the normal life history of cumulonimbus clouds occur. Thus, it is not unusual for the first rain stage to be interrupted as renewed convective activity carries the entire three-phase (ice, water, vapor) region above the critical temperature isotherm, to initiate another hail stage. Occasionally, too, a descending current may set in which will rapidly shrink the cloud by causing the evaporation of its ice or water elements, and cause the omission of one or several stages of the normal cloud.

The form of cumulus clouds usually gives an accurate indication of their stage in the life history pointed out above. The appearance of the ice cap, and its downward spreading into the main cloud mass, are especially clearly shown in nearly all cases. The ice portions of the cloud can always be detected by their indistinct edges, due to the trailing wisps of ice crystals, as contrasted with the sharp borders of the portions composed of liquid water droplets. It might be mentioned here that the "anvil-like" top of cumulonimbus clouds is not in itself any indication of the ice crystal nature of those portions of the clouds. Generally the critical temperature is attained far below the anvil, which itself only represents the spreading out of the ascending air currents as they strike a high level inversion. This inversion may even be the base of the stratosphere with large thunderstorms.

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CHAPTER 14

CLOUDS

INTRODUCTION

The study of clouds is one of the most popular phases of meteorology, for it is a subject in which everyone is more or less interested. From the most ancient times remarks on the shape and significance of clouds have been recorded. Until the dawn of the scientific era, however, no attempt was made to classify their various types or to associate them with weather phenomena.

Clouds have very definite scientific and practical uses to the meteorologist. Information concerning the direction and velocity of upper winds may be obtained by observing their motions. Useful weather indications often result from studying the development of certain cloud types. They assume a very important role in connection with forecasts made entirely from local indications.

CLASSIFICATION

The classification of clouds serves no useful purpose itself. In furnishing a means of studying other weather phenomena, however, it is of very considerable value. Lamarck introduced the first classification of cloud types in 1802. This was poorly received, no doubt because it was published in a journal which printed weather forecasts based on astrology. The following year Howard presented a classification in the Philosophical Magazine which was received with enthusiasm, and which is the basis of all modern classifications. It is interesting to note that Lamarck and Howard actually presented almost identical classifications, although they employed different names for the cloud types. Howard's three fundamental classes: cirrus, cumulus, and stratus are employed today with their original meaning.

The standard of present day cloud study is the "International Cloud Atlas," published by the International Meteorological Committee. This is a truly monumental work on the subject and includes unequalled illustrations of all cloud types. The classification adopted by the International Committee is the usual scientific one, including families, genera, species and varieties. Four families are introduced and each of these is divided into several genera.

These latter are then subdivided into a number of individual species. Due to the complexity of cloud types and the lack of any regularity of appearance, this form of classification is, of course, subject to considerable inaccuracy. It does provide a satisfactory means of discussing various cloud forms, however. One difficulty of this classification is that its major divisions are based on the *altitudes* at which the various clouds occur rather than on their *form.* This is not a very serious objection though, since with practice the observer will learn to correlate instinctively cloud heights with the various types observed.

Table 4 lists the various families, genera and species as given in the International Cloud Atlas of 1932. It is impossible to reproduce here more than a very few examples of the cloud types shown in this atlas and the reader is referred to it for illustrations of all of the various species.

TABLE 4

FAMILY A: High Clouds
(Mean lower level 20,000 feet)
1. Genus cirrus (form *b*)
2. Genus cirrocumulus (form *b*)
3. Genus cirrostratus (form *c*)

FAMILY B: Intermediate Clouds
(Mean upper level 20,000 feet, mean lower level 6000 feet)
4. Genus altocumulus (form *a* or *b*)
5. Genus altostratus (form *c*)

FAMILY C: Low Clouds
(Mean upper level 6000 feet, mean lower level close to the ground)
6. Genus stratocumulus (form *a* or *b*)
7. Genus stratus (form *c*)
8. Genus nimbostratus (form *c*)

FAMILY D: Clouds with vertical development
(Mean upper level that of the cirrus, mean lower level 1500 feet)

9. Genus cumulus (form *a*)
10. Genus cumulonimbus (form *a*)

Form a—Isolated, heap clouds with vertical development during their formation, and a spreading out when they are dissolving.

Form b—Sheet clouds which are divided into filaments, or rounded masses, and which are often stable or in process of disintegration.

Form c—More or less continuous cloud sheets, often in process of formation or growth.

HIGH CLOUDS

Cirrus clouds occur in extremely varied forms. They are divided into three main genera, the true *cirrus* clouds, the *cirrocumulus*, and the *cirrostratus*. All are composed of ice crystals.

FIGURE 94.—CIRRUS CLOUDS (*Cirrus uncinus* FROM INTERNATIONAL CLOUD ATLAS)

The banded arrangement of this cloud type is an indication of approaching warm front activity.

Cirrus clouds appear as delicate detached clouds without noticeable shading (figure 94). They are generally white in color and have a fibrous or silky appearance. They may occur as isolated tufts or

plumes, or may be arranged in curved lines or bands. These bands cross the sky and appear to converge on the horizon, due to perspective. In general cirrus clouds are so transparent that they do not obscure the sun's disc, and they are notable for their intense whiteness and silky edges. Occasionally halos are observed in cirrus clouds, although this is rather uncommon because of their great transparency. Cirrus clouds always occur at very high altitudes, generally ranging from 20,000 to 40,000 feet above sea level. In the higher latitudes they may occur at somewhat lower levels and

FIGURE 95.—CIRROCUMULUS CLOUDS (*Cirrocumulus undulatus* FROM INTERNATIONAL CLOUD ATLAS)

in tropical regions they may occur somewhat higher. In general they occur at lower elevations in the winter than in the summer.

Cirrocumulus clouds consist of a layer or a large patch, of small, white flakes, or very small round masses without shadows, which are arranged in lines or in ripples (figure 95). This type of cloud generally results from a transformation of cirrus or cirrostratus clouds. It should not be confused with altocumulus clouds. In general, it should not be identified unless its association with cirrus or cirrostratus clouds is evident. Cirrocumulus clouds occasionally exhibit halos—a characteristic of ice crystal clouds.

Cirrostratus clouds appear as a thin veil, covering much of the sky (figure 96). The sun or moon may be clearly distinguished through the cloud layer. Halos are very common. The cirrostratus veil may be so diffuse that it only gives the sky a milky appearance. At other times it may be sufficiently thick so that it occurs as a

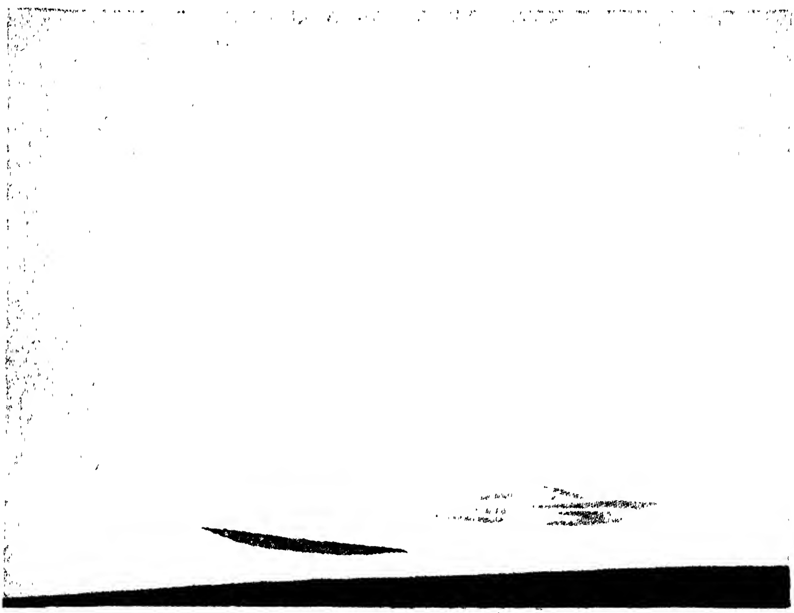

FIGURE 96.—CIRROSTRATUS CLOUDS (FROM INTERNATIONAL CLOUD ATLAS)
Note halo around the sun.

definite white veil covering the entire sky. It is never so thick that it blurs shadows of objects on the ground, and it is rarely gray or dark colored, as is the case with lower clouds. Frequently, a fibrous structure appears with cirrostratus clouds which have been derived from fibrous forms of cirrus clouds.

INTERMEDIATE CLOUDS

Altocumulus and altostratus clouds may occur through a wide range of elevations from approximately 6000 to 20,000 feet. They tend to occur at lower levels during the winter and they also occur

nearer the surface in the higher latitudes. Altocumulus clouds are composed of water droplets, while altostratus clouds are usually composed of ice crystals. They may be derived from the spreading out of the tops of cumulus clouds or through general frontal activity. They frequently occur in connection with other cloud types. It is common to observe two or three levels of altostratus or altocumulus clouds in a given situation. Frequently the altocumulus and altostratus clouds merge from one into the other within a single cloud sheet.

Altocumulus clouds appear as a layer, or as a series of patches or rather flattened, rounded masses (figure 97). These are generally

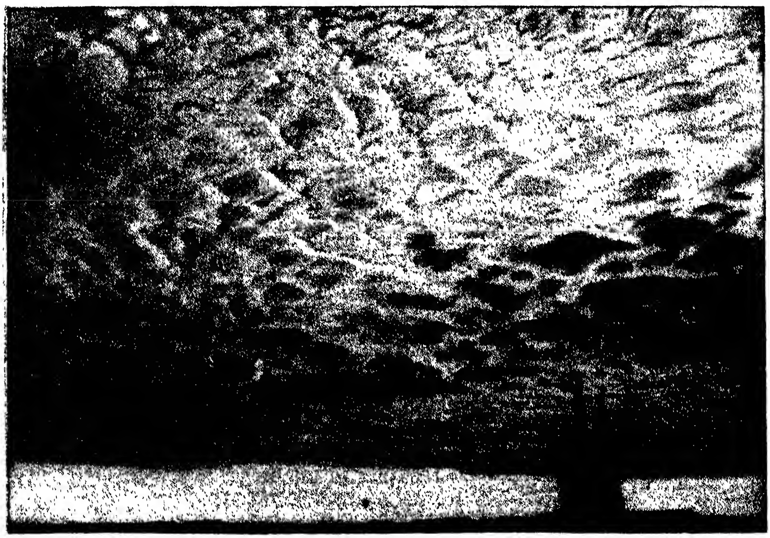

FIGURE 97.—ALTOCUMULUS CLOUDS (*Altocumulus translucidus* FROM INTERNATIONAL CLOUD ATLAS)

Some tendency toward a regular structure in two directions is displayed.

arranged in groups of waves in one or two directions. The individual cloudlets may be so close together that their edges join to form a single cloud sheet. Altocumulus clouds may or may not show shading. Usually they do exhibit more or less shading, particularly in cases where the individual cloud elements are well detached. When the cloud sheet is continuous over most of the layer, its characteristic cumulus form is indicated by a general irregularity shown by varying densities throughout the layer. This type of altocumulus is

easily confused with the altostratus types. Actually it represents a transitional phase from one type to the other. A slight change in the instability of the air may cause a change from one type to the other. High altocumulus clouds may be distinguished from cirrocumulus by the lack of unquestioned connection with cirrus clouds. They do not show the halos characteristic of the ice crystal types.

In the lower levels altocumulus may be confused with stratocumulus. They may generally be distinguished from the latter, how-

FIGURE 98.—ALTOCUMULUS CASTELLATUS CLOUDS (FROM INTERNATIONAL CLOUD ATLAS)

Of particular interest here are the little turrets, arranged in lines and resting on a common horizontal base, as in the lower third of the picture. Pronounced cumulus development is seen in some of the clouds in the upper part of the picture. This cloud type is a frequent precursor of thunderstorm activity.

ever, by the presence of colored rings or *coronas.* These coronas generally appear when a thin semi-transparent patch of altocumulus passes in front of the sun or moon. The colors are arranged with red outside and green inside and they may be repeated several times.

Several varieties of altocumulus are of considerable importance in that they are indicative of the unstable air conditions preceding thunderstorm activity. The form *altocumulus castellatus* frequently

occurs early in the day when conditions are favorable for strong convection. This cloud type consists of cumuliform masses with more or less vertical development, generally arranged in a line and resting on a common horizontal base. This gives rise to a very characteristic crenellated appearance (figure 98).

Altostratus clouds appear as a fibrous film, more or less gray in color, through which the sun or moon may appear vaguely as though seen through ground glass (figure 99). The sky generally has a milky appearance when the cloud deck is thin or it may be dark and

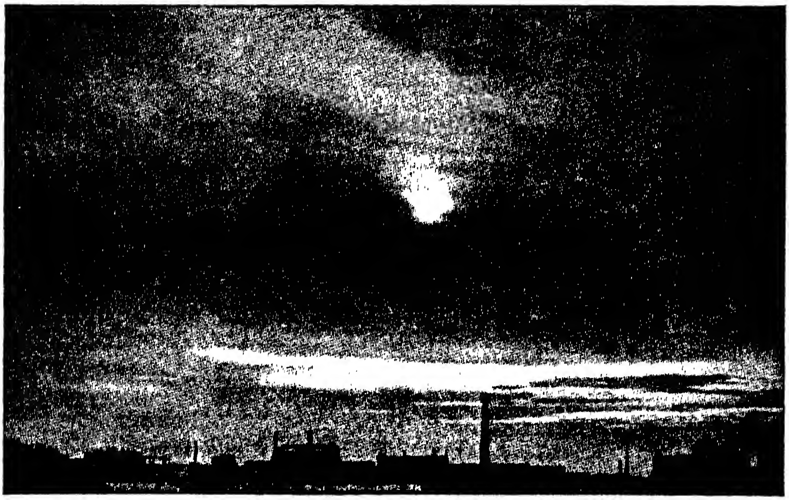

FIGURE 99.—ALTOSTRATUS CLOUDS (*Altostratus opacus* FROM INTERNATIONAL CLOUD ATLAS)

This cloud type often just precedes, and at times accompanies, active precipitation. It gives the sky a dark, forbidding appearance.

threatening when the cloud layer is thick and low. Precipitation in the form of rain or snow may fall from altostratus clouds. This type of cloud occurs throughout a rather wide range of elevations, from 6000 to 15,000 feet. It generally occurs at somewhat higher elevations than the altocumulus clouds. Altostratus clouds may be distinguished from cirrostratus clouds by the absence of halos and by the fact that shadows of objects on the ground are generally obscured when the sky is covered with altostratus clouds. They may be distinguished from the lower nimbostratus clouds in that the latter are consistently much darker and conceal the sun and moon in

all portions of the cloud, while the altostratus only hides them in the darker portions. Frequently trailing precipitation masses or *virga* occur within altostratus, occasionally reaching the ground to produce light rain.

Altostratus cloud decks frequently result from the transformation of a deck of altocumulus clouds. Conversely they may break up under convective activity to form altocumulus. In all cases the fibrous structure of altostratus distinguishes it from the entirely formless nimbostratus, and from the cumuliform appearance of altocumulus. A number of subdivisions of this general cloud type have been recognized and it will be necessary to refer to a cloud atlas, such as the International Atlas, to distinguish all of these types.

LOW CLOUDS

Three general types of lower clouds are encountered, all of which occur at levels below approximately 6000 feet. In many cases the distinction between clouds of this general class and that of the intermediate clouds is entirely a matter of elevation. Thus, a cloud layer which appears to be altocumulus to an observer near sea level may appear to be a stratocumulus cloud layer to an observer at an elevation of 5000–6000 feet and an altostratus deck viewed from a valley floor may appear to be of the stratus type to an observer on an adjacent high mountain.

Stratocumulus clouds commonly appear as a layer, or a series of patches composed of rounded masses (figure 100). They usually show considerable shading and may range in color from white to dark gray with a general soft appearance to the whole mass. The individual cloud elements are arranged in lines or waves in one or two directions. The various wavy lines may be so close together that they form a continuous covering to the sky, or they may appear as individual rolls separated by clear sky. Stratocumulus clouds generally show no close relationship with middle or high clouds although they may occur with them. They are often related to clouds of the general cumulus type, however. Stratocumulus clouds may pass into stratus clouds as general convective activity decreases, and vice versa. The principal distinguishing characteristic between the two cloud types is the wavy or rolled appearance of the stratocumulus as contrasted to the flat bases and tops of the stratus clouds.

Very frequently stratocumulus clouds cover the entire sky during the winter at an altitude of 1000 to 2000 feet above the surface. This particular occurrence of the stratocumulus type generally forms just beneath a marked temperature inversion and is usually associated with strong surface winds. The upper surface of a stratocumulus deck has a wavy appearance which frequently closely resembles the surface of the sea. Stratocumulus is distinguished from altocumulus by the coronas which occur with altocumulus but

FIGURE 100.—STRATOCUMULUS CLOUDS BELOW WITH IRREGULAR CIRRUS CLOUDS ABOVE PLANE

not with stratocumulus, and by the apparent larger size of the individual elements of stratocumulus clouds.

Stratus is a uniform cloud layer or veil giving to the sky a characteristic hazy appearance. It generally exhibits uniform upper and lower surfaces. It may occur with any other type of cloud. It frequently forms during or after the ceasing of precipitation, as *scud*. In general no precipitation falls directly from a stratus cloud, except perhaps a fine drizzle. In cases of active precipitation the low clouds producing them are of the nimbostratus type. When stratus clouds rest on the ground they form *fog*. Where the stratus cloud is broken the resulting cloud layer is called *fractostratus*. In cases where the cloud patches produced by the breaking up of a

stratus deck take on a cumulus appearance due to convection, the resulting broken cloud layer is called *fractocumulus.*

Nimbostratus is the low, formless, dark gray cloud layer from which precipitation in the form of rain or snow frequently falls. Even though active precipitation does not occur the cloud may be called nimbostratus if it fulfills the above definition. Occasionally nimbostratus may form from the gradual transformation of a stratocumulus deck. Nimbostratus is distinguished from stratus by the type of precipitation. This never amounts to more than a slight drizzle, in the case of stratus, while it may be moderate to heavy with nimbostratus. Nimbostratus clouds appear to be feebly illuminated from within, while stratus clouds show contrasts between lighter and darker portions and appear definitely to be illuminated from above.

CLOUDS WITH VERTICAL DEVELOPMENT

The general cumulus type clouds, which include the true cumulus and the cumulonimbus, all have marked vertical development. They may occur throughout a wide range of elevations from about 1500 feet to the upper cirrus level at the base of the stratosphere. This general class of clouds is readily distinguished by its marked vertical development, and by the dome-shaped appearance of the upper surface of the clouds, as contrasted with the very limited vertical development and flat appearance of the upper surfaces of other cloud types.

Cumulus clouds are marked by definite horizontal bases corresponding to the *saturation level,* and by considerable vertical development which, however, does not reach the *ice crystal stage.* The tops of true cumulus clouds are rather definitely limited, and the surfaces of the clouds generally appear well formed and clear cut (figure 101). Cumulus clouds generally develop on days when high clouds are lacking. They are usually the result of diurnal convective activity. Because of this, they generally first make their appearance in the morning, grow during the afternoon and disappear again during the evening. Cumulus clouds may disintegrate and form a layer of altocumulus as the general convective activity decreases. Conversely, cumulus clouds may be generated from certain types of altocumulus, especially *altocumulus castellatus.* Cumulus clouds exhibit

strong contrasts of light and shadow, appearing white on the sides toward the sun, and dark gray or almost black on the opposite sides.

There are two chief species of cumulus clouds. *Cumulus humilis* shows very little vertical development and is generally somewhat flattened. It is the characteristic cloud of fine weather. *Cumulus congestus* is the well developed cumulus cloud, with a cauliflower appearance. This frequently passes into the cumulonimbus type.

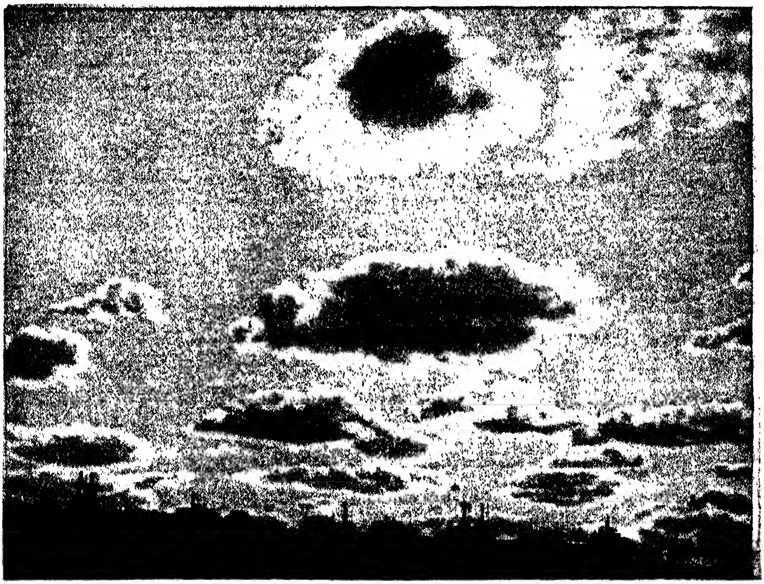

FIGURE 101.—FINE WEATHER CUMULUS (*Cumulus humilis* FROM INTERNATIONAL CLOUD ATLAS)

Cumulonimbus is a very heavy cloud with unusually marked vertical development (figure 102). Its summits rise in the form of great towers, and its upper portions have a fibrous texture often spreading out in the shape of an anvil. This cloud is characteristic of thunderstorm conditions and produces precipitation types indicative of strong convection, such as hail and graupel. If it is not possible to see enough of a cumulus type cloud to distinguish its genus, the presence of a brisk shower, or of hail or graupel, are sufficient criteria to characterize definitely the cloud as a cumulonimbus. The lower portions of cumulonimbus clouds generally resemble nimbostratus and frequently exhibit clearly defined virga.

The essential difference between the true cumulus and the cumulonimbus lies in the structure of the upper portions of the cloud. During intense convection the upper portions of cumulus clouds reach the freezing level. After the *ice crystal stage* has been reached, conditions are favorable for the production of heavy precipitation. The ice crystal stage is indicated by the presence of a fibrous veil around the upper portions of the cloud. In high latitudes, when the ice crystal level is near the surface of the ground, this

FIGURE 102.—CUMULONIMBUS CLOUD

Note the "anvil-shaped" top with mantle. This is a sign that the upper portions of the cloud have reached the *ice-crystal stage*. The uniform base lies at the *condensation level* for the region.

fibrous structure may extend through the entire cloud mass, so that the cumulus parts of the cloud disappear and are replaced by a mass of cirrus.

Cumulonimbus is the origin of many of the clouds which appear at the rear of large storm areas. The higher portions of cumulonimbus may be blown off to form a mass of dense cirrus, while the lower portions may produce thick sheets of altocumulus or stratocumulus. At times when the upper winds are unusually strong the upper portions of cumulonimbus may be blown out ahead of the general storm area to form a deck of dense cirrus some distance

in advance of the main storm. This type of cirrus is called *false cirrus* and does not show the delicate, feathery texture of true cirrus.

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1. International Atlas of Clouds and of States of the Sky. I. General Atlas (also published in abridged form). II. Atlas of Tropical Clouds. III. Atlas of Special Clouds. Office National Météorologique, Paris.

This is a truly monumental work on clouds and sky conditions. Every meteorological student should have access to the General Atlas, and every weather observer should have a copy of the abridged form of the General Atlas for constant reference. The Atlas consists of a collection of the finest obtainable cloud photographs, with numerous explanatory sketches, and a rather complete accompanying text.

2. Cloud Forms. U. S. Weather Bureau, Washington, D. C.

This is a collection of 12 half-tones of the more common cloud types.

3. Cloud Forms. H. M. Stationery Office, London, England.

This is a booklet containing definitions and descriptions, with photographs of the more common cloud types.

CHAPTER 15

FOG AND STRATUS CLOUDS

INTRODUCTION

The general subject of fog is perhaps the most important problem confronting the aeronautical meteorologist today. Fog is an obstacle that is present in one form or another in nearly every section of the earth except the tropics, where it is rarely found. It is a serious menace to the experienced pilot as well as the novice. It is often extremely difficult to forecast since its formation is usually intimately connected with very local influences. Until a satisfactory method is found to dissipate fog over landing fields the meteorologist must endeavor constantly to improve forecasting technique so that the hazards of fog to aeronautics may be reduced to the minimum.

Considering the importance of fog it is rather strange that so little has been written on its fundamental aspects. Perhaps the most satisfactory approach to the subject has been that commenced by Calwegen and carried on by Willett after Calwegen's untimely death, in which fogs are classified according to their relation to frontal systems and to air masses. Since the basis of the meteorological chart is the air mass, this classification is a fundamental one that lends itself readily to practical application and to subdivision and modification to suit special cases. Many other investigators have written on the general subject of fogs, including Georgii, G. I. Taylor —and their work has furnished much valuable observational data, but until very recently fogs were differentiated only as advection and radiation types, with the result that any logical approach to the subject has been impossible.

Today, in spite of the fact that much more is known concerning the situations under which fogs may form than was the case a few years ago, the detailed forecasting of fog is a most difficult task requiring intensive study and a thorough knowledge of local peculiarities of geography and topography. This is particularly true in doubtful cases where, for instance, slight air drainage may be

sufficient to prevent fog formation or where a slight shift in the wind direction may cause fog to cover an airport.

CLASSIFICATION OF FOG AND STRATUS CLOUDS

Several classifications of fog and stratus clouds are in common use among meteorologists. The one introduced by Willett in 1928 along modern lines, using air mass and frontal ideas, is perhaps the most generally satisfactory one available and will be used in this discussion with a few minor modifications. J. J. George has recently introduced a slightly different classification, as a result of his extensive work on fogs of the eastern and southern United States, although he has employed most of the terms of Willett's classification with only minor modifications. The chief aim of his classification has been to eliminate an overlap between certain types of advection and radiation fogs.

CLASSIFICATION OF FOG AND STRATUS CLOUDS

A. Air Mass Fog
- I. Advection Fog
    - a. Types due to the transport of warm air over a cold surface.
        1. Monsoon Fog
        2. Sea Fog
        3. Tropical Air Fog
    - b. Types due to the transport of cold air over a warm water surface.
        1. Arctic Sea Smoke
        2. Lake and River Steam Fog.
- II. Radiation Fog
    - a. Ground Fog
    - b. High Fog
- III. Maritime Fog

B. Frontal Fog
- I. Prefrontal Fog
    - a. Preceding a warm front
    - b. Preceding a cold front
    - c. Preceding an occlusion

II. Front Passage Fog
  a. Accompanying a warm front
  b. Accompanying a cold front
  c. Accompanying an occlusion
  d. Accompanying an upper front

III. Postfrontal Fog
  a. Following a warm front
  b. Following a cold front
  c. Following an occlusion

ADVECTION FOGS

Monsoon Fog—This is an advection fog, essentially a coastal type, is most prevalent during the warm season, and generally occurs in modified Polar Continental air. It is generally present during periods at which a marked temperature contrast exists between adjacent land and water surfaces, when the air exhibits a relatively high specific humidity, and when the general cyclonic activity is weak enough so that the monsoon circulation proceeds normally. The forecasting of monsoon fog along coastal regions depends upon the presence of the conditions mentioned above, and it generally may be expected therefore in regions in which upwelling of cold water along a coastal region is prevalent, so that a wide temperature difference exists between the cold waters and the warm coast. This limits its occurrence to those portions of the earth which show this phenomenon, examples of which are the New England coast and the northern European coast. Conversely, this type of fog may never be expected in the lower latitudes where the contrast between land and ocean temperatures is relatively slight. Since a moderately high specific humidity is also required, monsoon fog is very rare at the higher latitudes where the water vapor content is relatively low. Since the general cyclonic activity must be relatively weak in order to produce marked monsoon fog conditions, it is apparent that the most favorable localities for the regular occurrence of fogs of this type are either regions of slight meteorological activity (latitudes 20°–30°), or in regions of decaying cyclones.

Generally monsoon fog forms over the ocean within a few miles of the seacoast when relatively warm and moist air from the

land passes seaward in the upper branch of the monsoon circulation (figure 103). It may either remain at sea or be brought onshore with the diurnal sea breeze. Rarely does it extend far inland although it may extend well out to sea under pronounced monsoon conditions. If the general monsoon circulation is interrupted by a weak cyclone, the fog which has formed over the sea may be carried onshore, and in continental regions which are exposed to the sea it may move inland for a considerable distance. Willett mentions a case in which monsoon fog which had formed off the European coast moved inland over Europe for several hundred miles when a weak cyclone caused an onshore gradient, even reaching the Alps as a broad tongue of fog.

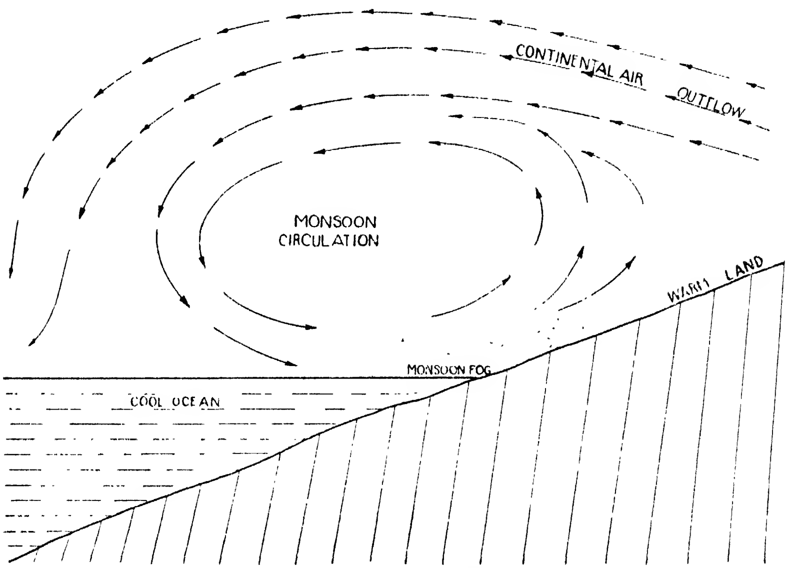

FIGURE 103.—PRODUCTION OF MONSOON FOG

The monsoon circulation along the coastal region carries warm continental air out over the ocean in the upper branch of the circulation. This air descends to the ocean, is cooled and carried back over the shore. Fog is formed in cases when the cooling is sufficient to reach the dew point.

Although conditions along the central California coast favor the occurrence of monsoon fog in this region, with the marked difference in land and ocean temperatures and the weak cyclonic activity, nevertheless the absence of monsoon circulation in this region, as evidenced by the lack of a land breeze at night or by an upper

branch of the circulation, causes the classification of the summer fogs of this region as *maritime fogs,* or as a mixture of *sea fog* and *maritime fog.*

Sea Fog—Whenever air over the open ocean is cooled by passage from warm to cold water so that its dew point is reached, a type of fog will be formed which is perhaps the most widely distributed of all fog types. Conditions necessary for its formation are: 1, a considerable contrast in temperature of adjacent parts of the ocean, and 2, moderate winds blowing from the warm to the cold portions of the ocean. This type of fog generally occurs in transitional types of Polar air during the intervals between periods of marked cyclonic activity. It is most common in the middle latitudes (35°–60°) where strong temperature contrasts over the ocean are frequent. It is especially common within regions occupied by cold ocean currents such as the Labrador current. In these cases old Polar air which has been warmed by passage over the general ocean surface and whose surface layers have attained a high degree of humidity, may reach saturation within a very short distance as it passes over one of the cold currents. The very marked foggy weather which is characteristic of the Newfoundland banks is caused by the air from the south passing northward over the cold Labrador current. In the far north, the moisture content of the air is generally so low that this type of fog is rare.

A very sharp boundary is frequently noted along the borders of cold ocean currents when sea fog is formed. This is particularly noticeable along the western border of the Gulf Stream as it moves northeastward off the southeastern coast of the United States. Here it is frequently noticed that the clear weather which prevails over the region occupied by the warm waters of the Gulf Stream changes to dense fog within a very few miles toward the shore, as the warm air which has been flowing over the warm regions of the South Atlantic encounters the relatively cold waters lying off the eastern coast of the United States. The thickness and extent of sea fog as well as the ceiling present underneath are greatly dependent on the past history of the air which is involved, as well as the wind velocity and temperature contrast.

In the early stages of formation over the open ocean, sea fog occurs as a series of bands, the spacing of which is related to the wind velocity. In the later stages, as the fog becomes more general, these bands coalesce and form a general solid fog bank which may,

however, even then, be marked by bands of denser and relatively lighter fog.

Tropical Air Fog—In the general widespread displacement of a tropical air mass from lower to higher latitudes, the cooling produced in the lower levels of the air as it passes over successively cooler and cooler water surface, frequently gives rise to a thick and widespread type of fog which is classified as *tropical air fog*. This differs from *sea fog* and *monsoon fog*, both of which are formed in air of comparatively recent polar origin which has been more or less modified by passage over warm waters. It differs from monsoon fog too in that it is related to general poleward movements of large bodies of tropical air whereas the monsoon fog is generally produced along coastal regions under the influence of relatively local circulations.

Generally, tropical air masses leave their source regions in the belt of subtropical high pressure fairly stable and rather dry throughout. As they move northward this stability is increased and inversions tend to form in the lower portions of the air. Cooling from contact with the cool surface below and radiation from the upper portions of the air mass cause a gradual increase in relative humidity until saturation is reached.

Fog produced by this type of air is generally of very wide extent since it is related to the entire air mass. Clouds are of the stratus type due to the inherent stability of the air mass. Often a heavy mist or drizzle accompanies this fog, giving rise to conditions of very poor visibility. It is interesting to note that tropical air fog is one of the very few types that may occur with strong winds. This is due to the inherent stability of the air and its increasing stability as it moves over colder and colder surfaces.

The cooling effect is naturally most pronounced in the forward portion of the air mass for here the surface has not yet been heated by the warm air and furthermore remnants of cold air from a former polar outbreak may mix with the warm current and accelerate its cooling. The most intense tropical air fog conditions, therefore, occur immediately behind the warm front making the first appearance of the tropical air and it is under these conditions that drizzle and mist are most frequently observed.

Only rarely does tropical air fog extend its influence over continental regions since in those seasons of the year when the continents are colder than the oceans the normal monsoon circulation

inhibits air movement from the sea to the land. It is only in cases of strong cyclonic activity in the winter that tropical air fog appears over continental regions. Over the ocean, tropical air fog may occur at all seasons of the year since a poleward temperature gradient exists throughout the year, but it is much more frequent in the winter when the gradient is the strongest.

Tropical Air Haze—Tropical air masses, as they move away from their source regions, exhibit poor visibility and considerable haziness, even before any condensation occurs. This phenomenon which is probably due to the presence of very minute dust particles gives rise to a characteristic opalescence which frequently serves to identify these air masses in the absence of other criteria. It extends throughout all levels in spite of the marked stability of the air mass, indicating that it must have originated during the early stages of the air mass history. Willett suggests that, at least in Tropical Atlantic air masses, the source of the dust is the desert region of northern Africa and that as the air moves northward from its source region the larger dust particles settle out, leaving the very fine ones in suspension to cause the opalescence. This characteristic haze disappears as soon as condensation washes out the impurities, thus it is most prevalent in the summer when the supply of dust is the greatest and when the amount of condensation is least.

Tropical air haze is generally of comparatively little importance to aviation since it usually does not greatly lower the visibility. Its chief importance is in aiding the forecaster to recognize tropical air masses and thus indicating the phenomena that may be expected from them.

Arctic Sea Smoke—This is an interesting phenomenon that is occasionally seen over water surfaces when the air is very cold. It is of no practical importance to aviation since it is very local in its occurrence. It is generally noted in breaks in the surface ice over Arctic regions or over large inland lakes which have not become completely frozen. The conditions necessary for the formation of this type of fog are a very stable stratification of the air, usually associated with a marked surface inversion, and an extremely low temperature of the air near the surface.

Lake and River Steam Fogs—The steam fogs which occur on clear nights over inland lakes and rivers, especially in the late autumn before they are frozen, are due to the same conditions which produce Arctic sea smoke. This type of fog is relatively local in its

occurrence but it may have very practical effects on aviation due to the fact that many airports in the interior of the continents are located near lakes and rivers.

These steam fogs are usually not over 300 or 400 feet thick, but they may be very dense. They are of especial importance along the Mississippi and Ohio rivers where many of the principal airports are located very close to the river banks. The general conditions under which they may form are: 1, sufficiently high humidity as expressed by a fairly small separation of the temperature and dew point early in the evening, 2, comparatively stable air, 3, light winds but not calm conditions, 4, a river temperature higher than the temperature of the adjacent land surface.

All of these conditions must be evaluated in the light of considerable experience for each locality and for each synoptic situation. It can generally be established, for instance, that with a separation of dew point and temperature of more than a certain number of degrees early in the evening, no fog may be expected during the night. Comparatively fresh polar air is usually sufficiently unstable so that the conditions necessary for the formation of this type of fog are not attained. The balance between wind velocity and fog formation is generally a very delicate one and must be given much consideration. With winds in excess of about 6 miles per hour this fog is rarely seen since the mechanical turbulence will generally destroy the surface inversion and furthermore the air transport will be sufficiently rapid so that condensation cannot occur. With very light winds, below 2 miles per hour, the air movement is usually so sluggish that the fog is confined to the immediate vicinity of the water body and rarely affects adjacent regions. A rather delicate balance exists furthermore between air and water temperatures, for if the water is considerably warmer than the air the inversion existing in the cold air may be destroyed. Thus the difference in temperature between the cold air and the warm water surface is an important factor to be considered in forecasting.

All of these matters are obviously of extremely local influence and must, therefore, be studied for each individual case. The season of the year, of course, is of some value in determining the possibility of formation of this type of fog, for in the summer, when the land surface and the air above it remain warmer than the river temperatures throughout the night, this type of fog cannot be formed. Furthermore, in the winter, when the contrast between the

air temperature and the warm water surface is too great, the stability of the air may be destroyed and the formation of the fog greatly retarded. Thus the times of maximum occurrence of this type of fog are in the spring and fall.

To a very limited extent, fog of this character may form in the vicinity of small inland lakes and may be of some importance at airports located in their immediate vicinity.

This type of ground fog, as has been mentioned before, is generally very thin and it is usually closely confined to the water surfaces over which it forms. It may form at any time after cooling commences during the afternoon or evening and it is almost invariably dissipated shortly after sunrise. Its limited vertical extent is often noted by pilots flying over areas occupied by this type of fog, when the tops of high buildings frequently penetrate the fog layer. The visibility on the surface, however, in this condition may be practically zero, so that it presents a real hazard to aviation. The tendency of steam fog to follow rivers closely is strikingly demonstrated when flying at night over an area affected by it. Each river valley has its ribbon of fog following every turn and twist as precisely as the river itself, while the adjoining hillsides are perfectly clear. The puffs of smoke from a locomotive running along the banks of the river through dense fog may occasionally be seen from above, penetrating the thin fog layer.

RADIATION FOGS

Surface Inversion Ground Fog—The formation of fog of this type depends on several factors—1, a surface inversion must be present, 2, winds must be light but not completely calm, 3, moisture content of the air must be high enough to produce saturation at temperatures which may be attained during the night, 4, sufficient hygroscopic nuclei must be present, 5, the topography of the region must be such as to allow the fog to accumulate.

These requirements, all of which must be met before radiation ground fog of appreciable thickness will form, are readily determinable and serve very definitely to limit the situations under which it may form. A surface inversion occurs only in the case of stable air masses, so all situations involving unstable air preclude the formation of radiation ground fog. Other types of ground fog or low stratus

may form in unstable air, particularly in the case of convergent activity, but radiation fogs never occur. The wind velocity must not be over 6–10 miles per hour in order to prevent the destruction of the surface inversion and to prevent the carrying aloft of surface moisture by turbulence. Completely calm conditions, however, prevent the slight surface turbulence necessary to produce cooling within the lower several hundred feet of the atmosphere and the subsequent formation of a thick stratum of fog. Under completely calm conditions radiational cooling at the earth's surface can proceed upward only by molecular diffusion, a process that can affect only the lowermost 3 to 5 feet during a single night. The optimum wind velocity for radiation ground fog formation varies somewhat in different localities, depending as it does on the topography of the region, but in general amounts to 3–6 miles per hour.

Radiation ground fog may be expected only when the humidity is fairly high early in the evening. The actual values of *relative humidity* or *dew point depression* in the early evening necessary to produce fog during the night can be determined for each locality only through experience. In order to do this the rate of cooling under various situations must be evaluated approximately. If it is found in this way that the dew point may be reached as radiational cooling proceeds during the night, the possibility of fog formation is very high. Only in cases of perfectly calm conditions, where the turbulence is insufficient to cause cooling of the lower several hundred feet, or where the topography allows the cold air near the surface to drain away as it is formed, or in cases where the air is very pure and free from condensation nuclei, will fog fail to form when the air temperature reaches the dew point. A chart similar to G. I. Taylor's Fog Graph, may be found useful in estimating fog probability from the dew point depression early in the evening. More useful than this, however, is a series of cooling curves for various conditions of temperature, humidity and cloudiness. With such curves it is possible to forecast very accurately the rate of cooling under varying conditions.

The quantity of hygroscopic particles available for condensation nuclei can be estimated in most instances by the haziness that exists during the late afternoon and early evening. If the visibility at this time is reduced to less than 4 to 6 miles because of haze (other than dust haze, which provides practically no active condensation nuclei) it may be assumed that ample nuclei are present to

allow the formation of a dense fog. If the atmosphere is free from haze, any fog which forms will be shallow and will tend to occur in isolated patches. Often the quantity of hygroscopic nuclei in the atmosphere can be estimated from the appearance of beams of light. With a high concentration of nuclei, the light rays will be scattered and the beams will be plainly visible, while in pure air little scattering will occur and the beam will be practically invisible as it passes through the air.

In perfectly level country air movement occurs only in response to pressure gradients. Where topographic irregularity is present, however, gravity flow of cold air is an important process in concentrating fog-laden air in low places. The drainage of cold air at times of general stagnant air flow is most important in determining the localities which will be affected by radiation ground fog in rolling country. In making terminal forecasts the meteorologist should give this factor careful consideration.

In estimating the chances of radiation fog in critical (if not all) cases, the trajectory of the air involved should be given the closest scrutiny. Thus, air which reaches a region after a day spent under an overcast sky will be much more likely to yield radiation fog than the *same* air arriving after a day under a clear sky. In the first case the amount of radiational cooling necessary to bring the air to its dew point may be very much less than in the second case. Thus, the very key to a successful fog forecast often is the study of the trajectory of the air.

Upper Inversion High Fog—The conditions necessary for the production of a high radiation fog are essentially similar to those which produce a radiation ground fog, except that the temperature inversion must be located at some distance above the ground, and the surface wind velocity may be considerably higher, frequently amounting to 15–20 miles per hour. Other conditions regarding moisture content and the presence of sufficient nuclei of condensation are approximately the same in both types of fog. Polar maritime air which has stagnated for several days and acquired one or more low level inversions generally furnishes the moisture and temperature distribution necessary to the production of high radiation fogs. It is interesting to compare the climate of the interior of North America with that of Europe in connection with this type of fog. The interior of Europe is open to direct invasion from the North Atlantic by maritime air masses which, under the influence

of an anticyclone centered over central or eastern Europe, may give rise to very widespread high fog. On the other hand the interior of North America is closed to direct invasion of maritime air by the great mountain ranges of the western portion of the continent so that this type of fog is almost unknown in the central part of North America.

After an inversion has been established in a region of anticyclonic activity, it tends to be intensified by further subsidence effects and by radiation from its upper surface. Furthermore, the temperature gradient in the air below the inversion frequently becomes fairly steep and as a result, turbulence within the surface layers tends to carry dust particles and other pollution aloft to the inversion. Thus, the maximum density of such impurities is generally found at the base of the inversion. It is also at this point that fog first begins to form, since the *relative* humidity increases from the surface up to the inversion. This is due to the fact that the *specific* humidity is more or less constant throughout the lower levels due to turbulence, whereas the temperature decreases aloft at a fairly rapid rate.

The existence and location of upper inversions is very important in the case of high radiation fogs and some means generally must be provided to detect them if successful forecasting is to be done. Aerographic soundings furnish this information, but since most soundings are made in the early morning the information obtained in this manner is of little use the following night. Airplane pilots, however, can readily obtain temperature soundings on practically all scheduled flights, and the information obtained in this manner is entirely satisfactory for use in locating inversion levels.

The shape of the temperature-altitude curve is often very indicative of the type of stratus deck that may be expected. Figure 104 illustrates a number of characteristic curves obtained by pilots late in the afternoon or early in the evening after the inversion level has become well outlined. Soundings taken during the middle of the day are of use only in cases of exceptionally strong inversions that are not destroyed by surface heating.

In figure 104, A shows a sounding in fresh, unstable air that has developed no inversion. Fog is very unlikely in this case. In B the same air mass is shown after slight modification and the establishment of a weak surface inversion, with the upper portion of the air still exhibiting a steep lapse rate. Under these circumstances a

thin layer of ground fog may form if the humidity is high enough. It will dissipate very shortly after sunrise. After considerable modification, the air mass may exhibit the temperature curve shown in C. Here a very pronounced surface inversion is present, together with a thick layer of air that is practically isothermal. Under these conditions a dense ground fog, that may be very persistent, is likely to form.

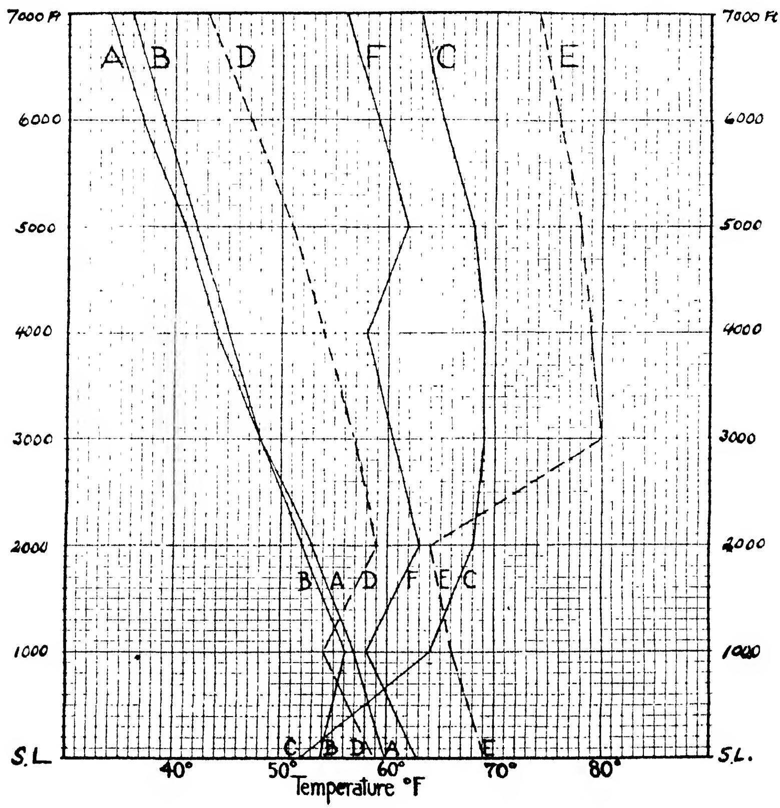

FIGURE 104.—TEMPERATURE SOUNDINGS IN VARIOUS TYPES OF AIR

D exhibits a case in which an upper inversion develops from a situation similar to B. In many cases, a surface inversion as in B will be forced aloft to form the upper inversion of D under conditions of moderate surface turbulence. A weak upper inversion such as this generally will yield only a thin cloud layer which will be dissipated soon after sunrise. On the other hand, a strong in-

version in connection with slight lapse rates near the surface and aloft as in E, may give rise to a very persistent cloud deck. Occasionally two upper inversions may become established as in F. In this case stratus decks may form at both inversion levels.

The information obtained from temperature soundings is of great use in determining the *top* of stratus decks, since the top will invariably fall very near to the inversion. The forecasting of the *base* of the cloud deck, however, is much more difficult. Theoretically this may be done if sufficient data are available regarding the distribution of temperature and humidity in the lower levels of the atmosphere. Practically, however, such data are generally unobtainable, and actually the balance between the temperature and dew point is so delicate in the cloud forming levels, that a forecast of ceiling height made in this manner is practically impossible. For practical purposes, the most reliable forecasts of ceiling heights may be made by considering the wind velocities in the lower levels. Almost invariably, a close relationship exists between the height of the ceiling and the wind velocity near the gradient level. With practically calm conditions, the ceiling is usually low, while with increasing velocities the ceiling tends to be higher and higher. A relation can thus generally be established for each locality between ceiling height and wind velocity which will prove to be very useful in cases of radiation fog.

HIGH RADIATION FOG IN CALIFORNIA

W. M. Lockhart has described an interesting occurrence of a particularly persistent type of high radiation fog characteristic of the Great Valley of interior California. The topographical situation of this region is particularly favorable to the production of intense and persistent inversions when the region is under the influence of the Great Basin HIGH. The entire Great Valley, from the vicinity of Redding on the north to Bakersfield on the south, has only one low level outlet to the sea—the Sacramento River Valley and San Francisco Bay. It is bordered on the east by the Sierra Nevada Mountains, which attain an average altitude of over 10,000 feet, and on the west by the Coast Ranges which attain an average elevation of several thousand feet. As a result of this geographic situation, maritime air reaches the Valley almost exclusively by

way of the Sacramento River Valley and to a very slight extent through a few low lying passes in the Coast Ranges. Under conditions of relatively stagnant airflow, practically the only means of ingress and egress from the valley is by way of the Sacramento River Valley.

During the winter when the Great Basin anticyclone shows marked development, with the formation of strong inversions, conditions become especially favorable for the formation of radiation fogs in the Great Valley. The first step in this process is the invasion of the entire valley by a mass of relatively fresh Polar Pacific air. Generally during the first two or three days after the appearance of this air mass, light fog will be observed in the Valley during the early morning hours and will be dissipated soon after sunrise. Within several days after the first appearance of fresh maritime air in the valley, the air mass of which this was a part, and which now occupies the entire western portion of the United States begins to undergo subsidence over the Great Basin region with the consequent outflow of warm dry air aloft from the interior over the coastal regions. This brings about the formation of intense inversions over the coastal valleys, with cold moist air entrapped below the valley walls, and warm dry air aloft. The turbulence within the air beneath the inversion is generally sufficient to keep it well mixed and as a result a stratus fog deck forms at the base of the inversion. With the stagnant airflow over the region, the air entrapped in the Valley undergoes little change and generally retains just sufficient turbulence to keep it well mixed. After the stratus deck has once been formed, conditions become increasingly favorable for its retention, since it acts as an excellent reflector by day and as a radiator by night, and this tends to intensify further the inversion. Under such conditions an exceptionally persistent thick stratus deck may cover the entire Great Valley for days at a time. During the night the lower side of the stratus deck will generally build downward, somewhat lower each night, until it reaches the surface to form dense ground fog. During the day, heating by the sun is very slight and the fog will lift slightly to form the ceiling of a few hundred feet in some localities, lowering again in the evening. The marked tendency of the fog to equalize surface temperatures is pointed out by Lockhart in one situation which he analyzed. Here the diurnal temperature range amounted to over 20° before the fog formed, whereas at the time when the

inversion was most strongly established and the stratus deck the most persistent, the temperature range amounted to but 6° F.

Once such a fog situation has become established it is frequently very persistent, and due to the low ceilings involved may hamper aviation operations very greatly. At times the entire Sacramento and San Joaquin Valley regions may be fogbound for several days at a time with ceilings everywhere less than three or four hundred feet. In the winter of 1934 a period of nearly three weeks occurred in which air transportation was practically halted throughout the entire Great Valley region.

The breaking up of this fog situation is caused generally by the approach of a fresh outbreak of Polar Pacific air which induces strong southerly gradients with a breaking up of the inversion and the replacing of the fog laden air in the valley. This is shortly followed by the appearance of the fresh unstable maritime air with excellent flying conditions. Although the situation described is somewhat unique, the general principles involved may be applied elsewhere with considerable value.

RADIATION FOG IN TROPICAL GULF AIR

An interesting means of forecasting stratus clouds which form during the summer in air of tropical maritime origin, has been worked out by J. J. George in connection with forecasting this type of fog over the southeastern United States. On investigating a number of cases of this type of fog, which is very prevalent over the entire southern part of the United States east of the Rocky Mountains, George found a rather close relationship between the time of fog formation and the wind velocity at the gradient level. He showed that with gradient wind velocities less than 15 miles per hour, fog never occurred. With velocities just above this critical value the stratus deck formed early in the evening, and as the wind velocity increased the time of its formation became later and later, until with velocities above 30 miles per hour, stratus fog never occurred. The formation of this type of fog is closely related to turbulence in the air mass in which it forms, and to the presence of a marked upper temperature inversion. As with all high inversion fogs, the top of the stratus deck can be fairly accurately forecast if the base of the inversion is known. The ceiling can also be de-

termined in this case if the surface condensation level is known, since the formation of the stratus deck depends on the carrying aloft by turbulence of air near the surface. Both of these factors can be determined in regions in which flying is carried on, if information regarding the temperature distribution of the upper air is obtained regularly during the late afternoon or evening. Since with the wind velocities mentioned above, the ceilings beneath the stratus deck generally exceed 300 to 400 feet, which is ample for airline operation over flat territory, the forecasting of actual ceiling height is not as important as an accurate forecast of the time of formation of the fog. Forecasts of actual ceiling height can be made by employing the gradient wind, since a close relationship exists between ceiling height and wind velocity as pointed out earlier.

The forecasting of this type of fog is occasionally made difficult by sudden changes in wind velocity during the forecast period caused by changes in the pressure field. This difficulty can generally be overcome by a study of the *pressure tendencies* at the time the forecast is made, so that any divergence or convergence in the isobars can be taken into consideration.

MARITIME FOG

Regions which are open to the invasion of comparatively fresh polar maritime air moving from a warm ocean surface over a cold land surface are often subject to *maritime fog*. The formation of this fog is caused by a combination of advection and radiation processes, with advection causing the movement of the fog producing air over the region affected, and radiation causing the principal cooling. In most cases this fog type will be found in winter, when the comparatively warm maritime air masses are cooled as they pass over the cold land surface. It is particularly common along the windward portions of continents that are little protected by topography from the prevailing wind. Thus, the west coast of Europe and the British Isles are very frequently affected by this fog.

In cases where the coast is bordered by a belt of cold water, a very close inter-relationship exists between *sea fog* and *maritime fog*. This holds true particularly along the west coast of the United States during the summer, when a very pronounced belt of cold

water lies off the California and Oregon coasts. Polar Pacific air masses which move onshore during this time of year are considerably cooled as they cross this cold water belt. They are then still further cooled during the night by radiation along the coastal region. That this type of fog has true sea fog characteristics is shown by the fact that it frequently lies offshore as a low stratus deck during the day and moves inland in the late afternoon and evening. That it is in part a radiation fog is indicated by the fact that widespread areas in the interior coastal valleys may become overcast simultaneously, as radiational cooling causes the air in these regions to reach saturation.

Conditions favorable for the formation of this type of fog are: 1, comparatively stagnant air flow over the general north Pacific region off the California coast, so that air which reaches the coastal region has had an opportunity to become somewhat stabilized, 2, an onshore gradient which will bring this modified Polar Pacific air into the coastal region from the west or southwest, 3, the presence of practically clear skies to afford ample opportunity for strong radiational cooling. Whenever these conditions are fulfilled, the formation of the typical high fog of this region is almost inevitable. Conversely, when any one of the above conditions is lacking, the formation of the fog is very unlikely. Thus, whenever the air which occupies this region is of comparatively recent polar origin, and is thus unstable in its lower levels, or when the general air flow is such that air reaches the coastal regions from the interior, rather than from the ocean, or when the sky is overcast with a high cloud deck, this type of fog is rarely formed. The problem of forecasting the ceiling beneath this fog and its time of appearance at coastal stations is especially difficult because of the disturbing effects of topography on local conditions. The top of the stratus deck can generally be foretold with considerable accuracy early in the evening, if the elevation of the temperature inversion is known and this is generally possible with the large number of airplane flights in this region. The forecasting of the ceiling under the fog, however, is a much more difficult matter. (The discussion of ceiling height forecasting on page 303 applies also to this problem.)

Several facts are of importance in forecasting this type of fog; 1, the elevation of the base of the inversion is generally very constant over widespread areas, so that in general the top of the fog forms a nearly level surface, 2, the elevation of the base of the

clouds is also generally very constant once the cloud deck has formed. These two factors greatly aid in forecasting the ceiling and thickness of this type of high fog since once it has formed over the immediate coastal area, it is possible to predict with confidence the conditions which will prevail when the cloud deck appears at inland localities.

Under fairly stable conditions with little cyclonic activity (as is normally the case during the late spring and summer months) the conditions which cause this type of fog change but little from day to day, so that the elevations of the base and the top of the cloud deck vary but slightly. It is generally found in this region that the stratus fog first appears as a very low stratus deck or a ground fog shortly after a fresh outbreak of Polar Pacific air. As the air mass becomes more and more stable the elevations of the inversion level and of the saturation level both increase, so that the ceiling and the top of the fog gradually raise day after day until three or four days after the initial appearance of the fog the ceiling amounts to 2000–3000 feet, after which the saturation level generally becomes so high that fog does not form. Generally, at about this time a new outbreak of Polar Pacific air makes its appearance and the entire process is repeated. It is thus seen that there is a marked periodicity both in the appearance of the fog and in its ceiling. This tendency for the fog to appear in cycles is related to movement of waves along the polar front which, at this season of the year, is located far to the north. It is generally observed that when the fog forms with low ceilings, it is dissipated early in the day and that as the ceilings increase the fog persists longer. Dissipation of the cloud deck occurs mainly from below, as the lower side of the clouds evaporates. This process proceeds rather slowly, generally at a rate of from 100 to 200 feet per hour until large breaks occur, after which dissipation occurs rapidly as the surface of the ground is heated and convective currents are established.

It is invariably found that the extent to which this type of fog invades the interior valleys of the coastal region is determined by the height of the inversion. With low inversions, the fog is confined to the immediate coastal areas and to the lower portions of valleys which open on to the coast. As the inversion level rises and the top and ceiling of the fog increase in elevation, it extends farther and farther into the interior. The elevation of the base of the cloud deck generally increases slowly toward the interior while the top

remains level, so that the clouds are usually thinner in inland localities than along the coast. For this reason the breaking up of the fog almost always occurs first in the interior, and then proceeds toward the coast. This phenomenon is also explained to some extent by the fact that dissipation of the cloud deck occurs most rapidly along its edges where heating of the ground surface is most pronounced.

Since the upper surface of the cloud acts as an excellent reflector, very little of the sun's radiation penetrates the clouds and as a result, heating of the earth's surface during the morning proceeds very slowly under the cloud deck. Ordinarily this increase in temperature beneath the clouds amounts to about one degree per hour until the cloud deck has begun to dissipate. It is generally found that dissipation of the clouds occurs rapidly in the lee of any hills across which the prevailing wind blows, due to the adiabatic heating of the air as it descends the lee slopes.

A type of fog that is intermediate between the typical *sea fog* and the *maritime fog* also occurs along the northern and middle Atlantic seaboard of the United States.

This type of fog is rather common in the spring months when, with weak cyclonic activity, a general southerly or southeasterly current of modified polar air reaches this region. The stratus decks formed in this manner are very important from the standpoint of aviation and are very difficult to forecast accurately. By the time an outbreak of polar air has passed out over the Atlantic ocean and has returned to the coast as a southerly current, it has generally become well stabilized and considerably warmed in its lower levels. As this air passes across the cold water lying on the continental shelf off the New England coast, the lower portions are considerably cooled and in many cases a sea fog of low stratus clouds is formed.

In many cases the balance existing between water temperature and air temperature is very delicately adjusted so that the actual formation of this sea fog may not take place until the air has almost reached the shore line. The matter of evaluating this balance is of prime importance in forecasting this type of fog and must be developed by a detailed study of surface winds, air temperature and ocean temperature. As a general rule it will be found that with wind velocity in excess of about 15 to 20 miles per hour, this type of fog will not form at sea due to the fact that the air is transported too rapidly to allow its lower portions to become cooled to the

saturation point. With lower wind velocities the fog will form at greater distances from the land. Frequently it will not form until radiational cooling over the land during the evening has cooled it below the condensation point. In this case it would be classified as *maritime fog* while when it forms over the sea it would be classified as *sea fog*. Actually all gradations between these types occur.

The ceiling under this fog is closely related to the wind velocity, since with very low velocities the fog may appear as a very low deck or as an actual ground fog, although ordinarily the wind velocity is sufficient to cause a ceiling of 500–1000 feet. The thickness of the fog layer is dependent largely upon the past history of the air mass. With a brief sojourn over southern waters only the surface layers of the air will have a high enough moisture content to produce fog, whereas if the air has been over the ocean surface for from five to ten days, the lower 1000–2000 feet may become saturated.

This type of fog rarely extends its influence inshore for a very great distance, being generally confined to regions which are essentially maritime. Over the open ocean as sea fog, it may occur at almost any time of year since the contrast between the various ocean currents is maintained at all times, although it is more frequent during the late spring and early summer when the temperature contrasts are the greatest.

Because of its mode of formation this fog may appear over a wide region in a relatively short time. Thus it is frequently noted that the entire southern New England coast from Boston to Baltimore will be affected by this fog within the space of a few hours. Its forecasting is therefore a matter of considerable importance. Since the air mass which produces the fog is generally rather uniform, it will be found almost invariably that the ceiling under the fog and its thickness will be very uniform over the entire area affected by it. In forecasting this fog it is very useful to have observation stations located along the immediate shore line, so that its first appearance along the shore may be noted, as it generally forms here before it does inland. In many cases this type of fog is present along the coastal waters during the day as sea fog, preceding its appearance over the land, and as the solar heating diminishes it tends to move inland. After it has become established during the night it generally persists until several hours after sun-

rise. In all but the most severe cases, it dissipates by mid-morning. The ceiling, which generally amounts to several hundred feet when the fog first appears, lowers during the night as the fog builds downward toward the ground, until in many cases a dense ground fog is present by midnight. In the morning the ceiling rises rather slowly, usually at a rate of about 100 feet per hour. The density of the fog near the surface depends to a large extent on the amount of atmospheric pollution, so that in the vicinity of large cities such as New York and Philadelphia, the visibility near the ground is invariably very poor.

This type of fog may occur along almost any coastal region bordered by a belt of cold water and subject to invasion by air of polar origin which has passed over a warm water surface. At times it appears to be *sea fog* and at other times typical *maritime fog.* Its forecasting depends to a great extent on a careful study of the local situation, and upon detailed analyses of past fog situations so that relations may be established between:

1. The wind velocity and, both the appearance or nonappearance of the fog, and the ceiling height,
2. The past history of the air mass (the number of days it has remained over warm water) and the thickness of the fog formed, as well as its appearance or nonappearance,
3. The contrast between surface water temperatures, and the temperature in the lower levels of the air.

Due to the irregularity of surface wind observations, it is preferable to use the wind velocity at either 500 or 1000 feet above the surface in evaluating (1) above. If upper wind observations are not available it may be found desirable to use the gradient wind as determined from the pressure field instead of the surface wind.

FRONTAL FOG

Convergent Fog—One of the most important types of fog and one which is directly related to frontal activity is that caused by convergence in the cold air beneath either a cold front or a warm

front, or an occlusion. From a reference to figure 105, it will be seen that any condition which causes a convergence of the cold air beneath a front toward the frontal surface will cause a forced ascent of the cold air beneath the front. If the humidity of the cold air is nearly at the saturation point before this ascent starts, it is clear that condensation will occur within a short time after the cold air commences rising. Since convergence caused in this way may cause the ascent of cold air over a very wide region under a frontal surface, and since the rate of ascent is frequently relatively uniform over a wide area, this type of fog may affect simultaneously a wide-

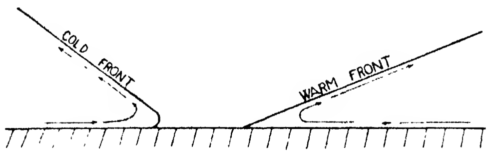

FIGURE 105.—CONVERGENT FOG FORMATION

The ascent of the cold air beneath either a cold front or a warm front causes it to coo adiabatically. As it becomes saturated, a cloud deck forms. Convergence is generally due to retardation of a front due to frontogenesis or cyclogenesis.

spread region. Because of this feature and because of the fact that the ceilings and visibilities under the cloud layer formed in this manner are frequently very low, the forecasting of this type of fog takes on particular significance for aviation.

The occurrence of convergent fog involves the presence of three factors:

(1) The presence of a fairly slow moving front which may be of the warm front, cold front or occluded front type,
(2) The presence of convergence toward the front within the cold air mass beneath it.
(3) The presence of sufficient humidity within the cold air to cause a relatively low saturation level.

The first and third of these conditions for the formation of convergent fog usually may be recognized without difficulty from an inspection of the weather chart. In this manner many synoptic situations can be seen at a glance to be incapable of yielding this type of cloudiness. If these two conditions are present however, the second criterion, that of convergence within the cold air mass,

must be carefully looked for and an attempt be made to determine its approximate magnitude. Perhaps the best method of determining the amount of convergence within the cold air is to construct an isallobaric field in the vicinity of the front. Any tendency toward convergent winds in the cold air will generally be shown in this manner. In the absence of information for constructing such a field, the wind directions and velocities within the cold air mass may be employed for the same purpose, since deviation of the air flow from the normal cyclonic flow is a measure of the amount of convergence or divergence.

Generally, it is possible to estimate with a fair degree of accuracy the elevation of the base of convergent cloud systems if the condensation level of the cold air is known. This may be determined approximately if the temperature and dew point of the surface air are known assuming that the rate of cooling caused by the vertical movement will amount to 5.5° F. per 1000 feet. Thus, if the temperature and dew point at the onset of convergence are 52° and 49° respectively, saturation of the surface air will occur after it has been lifted about 500 feet. An estimate of the time required to reach saturation may also be made if the rate of vertical lifting is known. This may frequently be calculated if the component of the wind normal to the front near the surface is known, together with the slope and velocity of the front. In figure 105, the wind velocity normal to the cold front is 7 miles per hour. The front is moving at the rate of 5 miles per hour. The *net* wind velocity normal to the front is thus 2 miles per hour. This will ascend the lower surface of the front at approximately $\frac{2}{50}$ miles per hour, if the slope of the front is $\frac{1}{50}$. This amounts to a vertical velocity of about 200 feet per hour, so if 500 feet of lift are required to reach saturation, 2½ hours will be necessary.

Certain general rules for the forecasting of this type of fog have been worked out, so that even in the absence of all of the specific information mentioned above, it is generally possible to forecast this fog with considerable success. In connection with cold fronts, it very rarely forms unless the front is being subjected to retardation. Thus, a cold front which is moving with uniform or accelerated velocity, practically never is followed by a convergent cloud system. On the other hand, one which is being retarded,

often due to the formation of one or more waves along it, is particularly subject to this phenomenon.

Furthermore, since convergent cloud systems result from the cooling due to adiabatic expansion of the cold air, it is clear that any other processes which aid or retard the general cooling will have an important affect on the cloud formation. Thus, it is invariably found that when convergent conditions develop in the late afternoon or evening, that radiational cooling greatly aids the adiabatic cooling process and in these cases the formation of the low cloud system may take place with unusual rapidity. On the other hand when the convergent process occurs during the morning, insolational heating greatly slows down the adiabatic cooling process so that in such cases the formation of the low cloud deck may be greatly retarded.

The formation of a convergent cloud system is often closely related to other fog producing agencies. One of these, radiational cooling, has already been mentioned. Others include the saturation of the air in advance of a warm front and following a cold front by the ordinary precipitation process; and the pressure drop which precedes a warm front and which produces a certain amount of cooling by expansion. The first of these two processes is of considerable importance in affecting the rate at which convergent cloud systems may form, since it affects the condensation level within the cold air. The second factor is, in most cases, of negligible importance since the magnitude of the pressure drop preceding a front generally is too small to produce a cooling of over 2° F. (a pressure drop of 0.2 inch in 3 hours, which is a fairly large value for the pressure change, will produce an expansion sufficient to lower the temperature only 1.2° F.).

Convergent cloud systems are of particular importance to aviation, not only because of the low ceilings and visibilities associated with them, but because a very large area may be affected within a very short time. It is not uncommon for regions of 10,000–30,000 square miles in advance of a warm front to become overcast by a low stratus deck, caused by convergence, within a space of two or three hours. With a cold front, the area affected may be somewhat less, but the speed of formation is often even greater than in the case of the warm front. Frequently the ceiling drops very rapidly after the first appearance of the convergent cloud system, until within two or three hours it may lower to

200–300 feet or even less. A fine mist or drizzle often accompanies these low ceilings, greatly reducing the visibility.

In the initial stages of the formation of convergent cloud systems, they rarely form near the intersection of the front with the ground. The reason for this will be made clear by referring to figure 105 from which it can be seen that the lifting of the cold air necessary to produce saturation is not sufficient in the area near the surface position of the front, whether it be a cold front or a warm front. Since the slope of cold fronts is generally much steeper than that of warm fronts, this area which is unaffected by convergent cloudiness is much smaller in the case of the cold fronts. This fog-free area generally disappears shortly after the onset of the convergent activity, however, as the air near the surface becomes more and more saturated with the mist or drizzle which accompany this situation.

The convergent cloud system associated with a warm front will generally cease abruptly at the warm front surface, so that the warm sector is ordinarily free from fog. In the case of convergence behind a cold front, the area of convergent cloudiness will usually move with the cold front, although once an area is affected by this type of cloudiness it will generally remain overcast for a considerable time.

Convergent fog, as may be seen from the above discussion may be either prefrontal, when it is associated with a warm front or an occlusion; postfrontal, when it occurs with a cold front on an occlusion; or front passage in the case of strong convergence with a cold front or occlusion which affects the cold air mass entirely up to the front. In the discussion to follow, only fog types whose formation is not directly related to convergence will be considered. Many of these situations will be intensified by convergent activity. In fact the convergence will in many cases be of considerably greater importance than the other fog producing agencies.

Prefrontal Fog Preceding a Warm Front—In this case fog is produced mainly through saturation of the lower levels of the cold air by precipitation from the overrunning warm air. The scud clouds formed in this manner may be very low, but they will also be extremely variable, being very thick and solid in areas of heavy precipitation and relatively thin and scattered where the precipita tion is less intense. The variable nature of these clouds makes accurate forecasting of ceiling heights practically impossible, since the

ceiling may vary from 200 feet to 4000 feet within the space of a few minutes as the scud clouds drift about.

Where the circulation induced in the cold air by the approach of a warm front causes its forced ascent over uneven topography, low stratus decks may form. The points at which these clouds may be expected can be determined from a study of the topography, and the likelihood of their formation can be determined from the degree of saturation of the cold air. In air which is nearly saturated by convergence, or by precipitation from an upper warm front, the additional lift caused by the ascent of low hills often causes this type of fog to form. This is particularly noticeable in the low rolling hills of Tennessee, Kentucky and Ohio, in the case of a warm front to the south. Several hours before the entire region is affected by convergence, the higher areas will be blanketed by low stratus clouds.

Prefrontal Fog Preceding a Cold Front—This fog type may be formed when the pressure drops preceding the front are very large, leading to adiabatic cooling of the air in the warm sector, or when the air in the warm sector is forced to ascend rolling topography by the general circulation. Neither of these processes occur, however, except in exceptional cases, and usually are of very minor importance. The ceilings under stratus cloud decks formed under these conditions are generally ample, except occasionally in the second case where the clouds may rest directly on the higher hills.

Prefrontal Fog Preceding an Occlusion—In the case of a cold front occlusion, the fog situation is very similar to that preceding a warm front. Saturation of the cool air beneath the warm front aloft by warm front precipitation may cause the formation of irregular scud clouds. The forced ascent of the cold air over higher topography, and the pressure drop preceding the frontal passage, may both contribute to adiabatic cooling of the cool air. Ceilings will be extremely variable as with the warm front stratus clouds.

Front Passage Fog Accompanying a Warm Front—A low cloud deck often accompanies the passage of a well developed warm front as scud clouds and stratocumulus clouds merge to form a cloud system moving with the front having ceilings of 200–1500 feet. Only rarely does the ceiling drop below 200 feet in level country, since the wind velocities are generally sufficient to keep the

cloud bases above this level. When the land surface is snow-covered, however, or over exceptionally cold ocean waters, true tropical air fog may form with the passage of the front. Considerable improvement generally occurs very shortly after the passage of the warm front, but the visibility may remain comparatively poor until the tropical air haze or fog is displaced by fresh polar air. When a warm front passes over mountainous regions it is frequently held back by the topography, so that it may not immediately pass at the surface. In such cases the improvement which followed its passage before reaching the mountains, may not occur over the higher topography, and continued prefrontal phenomenon will continue until a cold front appears and displaces the warm air.

Front Passage Fog Accompanying a Cold Front—The only fog in this case is the scud that often appears with the heavy showers as the front passes. Low convergent stratus may appear with the front, however, and should be expected whenever conditions are favorable for convergent activity.

Front Passage Fog Accompanying an Occlusion—In this situation the low clouds accompanying the frontal passage are essentially those of a cold front with scud clouds in the showers and generally clearing weather except when convergent activity occurs.

Front Passage Fog Accompanying an Upper Front—When an active upper front in connection with a warm front type occlusion passes over a region it is frequently accompanied by a belt of showers. Occasionally the ceilings in these showers may become low for a short time due to scud clouds. This is particularly true when considerable snowfall occurs, with the formation of a low stratus deck in the cold air near the surface. In warm weather, turbulence in the cold air is generally sufficient to give ample ceilings under these conditions.

Postfrontal Fog Following a Warm Front—In general, rather marked improvement in low cloudiness follows the passage of a warm front with the consequent cessation of convergent activity, unless the warm air is of true tropical origin. If it is a tropical air mass, mixing with the cold air being displaced near the warm front will intensify the tropical air fog normally to be expected from its northward movements. Thus, the forward edge of an advancing tropical air mass occasionally exhibits fog in a very aggravated form, especially if the earth's surface is especially cold.

Postfrontal fog following a cold front or an occlusion is of im-

portance only when convergence occurs. Otherwise the displacing cold air mass almost invariably brings greatly improved conditions.

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CHAPTER 16

SPECIAL PROBLEMS OF AVIATION

INTRODUCTION

Certain weather conditions are of interest almost exclusively to aviation. The problem of turbulent air, is of slight concern to farming or shipping interests, for instance, yet it is very important to the pilot of an airplane. Similarly, few persons in ordinary business are concerned with the motion of the upper winds, yet these are as important to the airplane pilot as the drift of ocean currents is to the sea captain. Some of these special problems of aviation will be considered here.

UPPER WINDS

Successful forecasting of upper winds is chiefly a problem of the accurate construction of free air pressure maps at flying levels. Since frictional forces are very slight in the free atmosphere, streamlines follow the isobars very closely, and an accurate pressure field will therefore indicate the wind directions very closely. These maps should, of course, indicate the locations of fronts at the level which the map represents, since the locations of these wind shifts are very important in preparing forecasts. The chief obstacle to the regular preparation of upper air charts at the present time is the lack of sufficient data. Upper air pressures are obtained from aerographic soundings, and wind directions and velocities are obtained from pilot balloon observations. Actually these sources of data fail frequently with present methods of obtaining them, particularly in inclement weather when they are most urgently needed. Development of the radio-meteorograph and radio directional method of determining upper winds offers the greatest promise toward increasing the accuracy and reliability of upper-wind observation and forecasting.

Even with the scanty data usually available, the meteorologist will find that the time spent in constructing one or more upper air

synoptic charts is extremely well spent. On these charts, which should include approximately the same area covered by the surface charts, are indicated the wind direction and velocity at the level for which the chart is constructed, using all available pilot balloon information. Pressures at the chart level also should be entered at all stations from which aerographic soundings are available. Positions of frontal surfaces at the chart level should then be indicated, using the surface weather chart as a guide to aid in locating them. The upper air pressure field then should be constructed, using the

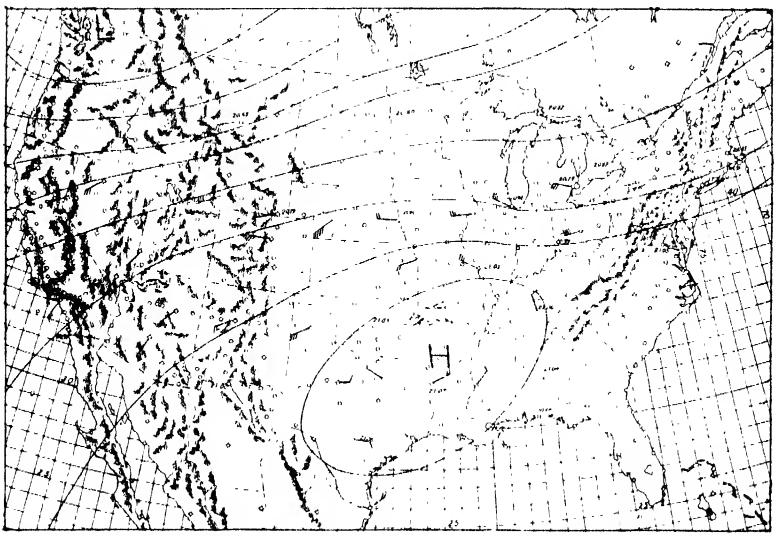


FIGURE 106.—CHART OF FREE AIR PRESSURE AT 10,000-FOOT LEVEL FOR THE UNITED STATES

Isobars are drawn for each 0.1-inch pressure. Pressures beside stations are given in inches. Wind arrows fly with the wind, each feather representing 10 miles per hour, and each half feather 5 miles per hour.

greatest possible care in drawing the isobars. Generally, isobars should be drawn for every tenth of an inch difference in pressure, and they must be carefully smoothed to conform to the velocity field and the synoptic situation (figure 106).

The wind velocity at various upper levels may be determined readily from the pressure field. Table 24 in the Appendix gives the wind velocity for various curvatures and spacings of sea level isobars. These values may be converted to upper air values by the use of table 23.

Even at times when pilot balloon data are almost entirely lacking over large areas due to overcast skies, the careful construction of the upper air pressure field may yield information of the greatest value with regard to the upper winds. Experience has shown that considerable confidence may be placed in upper wind forecasts prepared from carefully constructed upper air pressure maps.

In most instances, one upper air pressure chart at 10,000 feet will suffice to show the wind distribution at flying levels, since this indicates the winds in the region from 7000 to 13,000 feet with fair accuracy. The winds below the 7000 feet level generally do not differ greatly from the gradient winds. In case the lower winds (from 3000–7000 feet) deviate considerably from the gradient winds, it will be found desirable to construct another upper wind chart at the 5000 foot level. If flights are contemplated at altitudes above 12,000–13,000 feet it will also be found very desirable to construct a chart for the 15,000 foot level. The latter chart will indicate wind conditions up to about 18,000 feet which is above the maximum cruising altitude for airplanes with unsealed cabins.

Upper air charts need be constructed but once a day, since the surface synoptic charts may be taken as an entirely satisfactory guide to developments in the upper atmosphere during the day. These upper air charts serve the additional purpose, besides indicating upper wind direction and velocity, of clarifying many situations that are not clear from surface data alone. Their construction takes but a short time, since only a few data are available, and the results obtained are frequently most gratifying.

Attempts at forecasting upper winds from the surface weather situation are usually rather unsatisfactory. In general, winds at the higher levels are shifted from their normal directions and velocities less than surface winds, during storm activity, but the actual shift from normal values is directly dependent on the depth of the storms, so any generalizations in this regard are practically meaningless. It may be said that any attempt to forecast upper winds without actual upper air data in the form of pilot balloon observations and aerographic soundings is doomed to almost certain failure. From a statistical viewpoint, certain generalizations with regard to the diurnal change in wind velocity and direction are interesting, but in most cases the dynamic changes caused by variations in the pressure field and synoptic situation are of far greater importance, usually masking the diurnal changes completely.

UPPER WIND DETERMINATION FROM PLANES IN FLIGHT

From the standpoint of the airline pilot and meteorologist a vast amount of potential information with regard to upper wind directions and velocities is never obtained. Every flight which follows a definite course at a definite altitude should be able to determine the direction and velocity of upper winds with an accuracy absolutely unobtainable with pilot balloon equipment, yet very few pilots take the trouble to determine this information. The only calculation required is the solution of the wind and drift triangle. This can be accomplished graphically on any one of several well known air speed calculators.

The pilot needs only to know his true air speed, ground speed, and drift angle. The true air speed may be determined to within 2–3 miles per hour if the altitude and indicated air speed are known. The ground speed may be determined to within 2–8 miles per hour, depending on the accuracy of determining time and distance over the ground. If the pilot is flying a course marked by radio ranges so that the time over check points may be determined exactly, and if he uses a stop watch to obtain his time between check points, he should be able to determine his average ground speed to within 1–2 miles per hour. If the flight takes place over regions where accurate checks are not possible, the ground speed may be in error as much as 6–8 miles per hour. The drift angle should be determinable to within 2°–3°, if some care is taken in holding a precise course, or if an automatic pilot is used. (In either case the gyro-compass should be reset frequently.) If a radio beam is flown, considerably greater accuracy is generally obtainable than if visual flying is being done, unless, of course, the radio beam happens to be affected by marked swinging.

In any case the solution of the wind triangle using the data obtained with the accuracies indicated above will result in the determination of the wind at the flying level with an accuracy of from 2–8 miles per hour and within 5°–7°, both limits of error considerably better than are obtained in practice with pilot balloon observations. Certain precautions must be taken to insure this degree of accuracy:

1. The ground speed must be determined as accurately as possible. This requires that times and distances be

measured very carefully. The time over a given spot during contact flight should be determined by looking vertically downward, and not by making a rough estimate of the time of passage. When flying a radio range, the time of passing the "cone of silence," or the intersection with an "on-course" signal in a beam being crossed, may be employed with great advantage. The use of a stop watch greatly improves the quality of the work.

2. A single, unchanging course from one check point to another should be carefully held, especially in contact flying, even if this involves a slight error in crossing the check point being approached. The error in the course flown can be calculated after the new check point is reached, if the distance by which it was missed is known. The correct drift angle of the course just flown can then be calculated and any error in the original drift angle may thus be obviated. This would not be possible if the course were changed during flight. When flying a radio beam under changing wind conditions, it will not be possible to hold a constant heading, and thus the drift angle will undergo a continuous change. In such cases, unless a pronounced and sudden wind shift is encountered, the *average* drift angle for the interval between the two check points may be employed in determining the *average* wind, a procedure that is entirely satisfactory, since the average wind over a small region is actually of more practical value than several isolated values. The method indicated above for correcting the drift angle in contact flying also tends to yield an average wind. When a sudden change in wind velocity or direction is detected between check points, that particular problem cannot be solved unless the exact time and location of the wind shift is known.

The calculations necessary to determine the direction and velocity of the upper wind may actually be performed in a very short time—generally in from 2 to 3 minutes at the most. Since the pilot generally calculates his true air speed and ground speed at each check point in any case, the additional time required to calculate

the upper wind will be found to be very slight if the drift angle is known.

In practical use, the drift angle and true air speed may be calculated just before reaching a check point. Immediately after passing the check point, the ground speed may be calculated and the wind direction and velocity can then be determined in a minute or less. This information can then be transmitted by radio, at the time the passing of the check point is reported. When position reporting is done at certain *times*, rather than at certain *points*, the wind velocity and direction at the last check point may be reported on each regular radio schedule.

ROUGH AIR

Atmospheric turbulence, which may be caused in several ways, is often experienced by aviators in the form of bumps in the air. These bumps occur when the airplane passes from a region of a certain vertical velocity to a region with a different vertical velocity. As the vertical velocity of the airplane is changed in passing from one region to another an acceleration in vertical motion will be produced that is felt as a bump. (There are, of course, no such things as "air pockets" with their implication of voids in the air. The air is a liquid filling all space as much as the water of the ocean fills the entire ocean basins and it would, therefore, be as impossible to have a "pocket" in the air as it would be to have a void in the ocean.) The sensation produced by bumpiness in the air is identical with that produced by the sudden movement of an elevator, in that it affects the balance mechanism of the human body and creates a certain definite sensation. Since this mechanism of the body, which is located in the inner ear, is sensible chiefly to *accelerations* and not to vertical *velocities,* it is only the *change* in the vertical movement of the airplane that is noticeable. Pilots thus find that although they feel very distinctly the bump when encountering a vertical air current, that they feel no particular sensation other than an apparent increase or decrease of weight, after they have become adjusted to the new vertical velocity.

Since the rate of change of vertical movement produces the sensation of bumpiness, it is apparent that *the more rapidly an airplane flies through bumpy air, the more severe the bumps will be,*

since the vertical accelerations encountered in entering and leaving vertical currents of varying velocity will be greater than would be the case if the airplane were proceeding more slowly. If it is necessary to fly through bumpy air, it will always be found that the discomfort attendant upon this type of flying will be greatly decreased if the speed of the airplane is lessened. Also, of course, the strain on the airplane will be greatly decreased if the speed is lowered. Many pilots recognize this fact and reduce their speed when flying through thunderstorms or other situations where unusual bumpiness is encountered.

Turbulence may be of several different kinds, each one of which occurs under certain circumstances and produces certain results. The two main types of turbulence are, (1) those due to thermal effects and (2) those due to mechanical effects. The former class is most pronounced during the afternoon when solar heating is at a maximum, and the latter is generally observed during strong circulation and intense cyclonic activity.

Thermal Turbulence—This type is due to differences in temperature between adjoining portions of a body of air. If these temperature differences are distributed so as to cause a steep vertical lapse rate, relatively light air below will rise though denser air above to produce columns of ascending air. These usually will be balanced by similar columns of descending cold air elsewhere in the body of air. Conditions favorable for this type of turbulence are strong insolational heating and rough topography. When these conditions are combined, as on hot summer afternoons in the arid southwestern portion of the United States, the turbulence may be very marked indeed. Generally this roughness is confined to the lower 2000–4000 feet above the ground. Thermal turbulence is also manifested when the boundary between widely different types of surface is crossed. Thus a distinct turbulent zone is generally present along the borders of lakes or oceans and even between wooded and prairie lands. The remedy for thermal turbulence is to fly at an altitude of several thousand feet above the surface.

A different type of thermal turbulence, generally in the form of one distinct bump, is encountered in flying through a strong *inversion*. Here the lift on the airfoil changes suddenly and causes a bump, generally rather slight, as the inversion is penetrated. In cases of an exceptionally strong inversion a short distance above the surface, the loss of lift involved in suddenly entering less dense

air on the takeoff, may cause some confusion to the pilot, especially, if he is climbing near the stalling speed. In all cases where a strong inversion is suspected near the surface it is well to attain ample flying speed early in the climb. Airships frequently find it difficult, it not impossible, to effect a landing when a strong low level inversion is present, since the excess lift exerted by the cold, dense air near the surface may make it impossible to penetrate the inversion.

Cumulus Turbulence—A special form of thermal turbulence is that due to the release of conditional instability in saturated air. In cumulus clouds, an initial upward impetus to air particles provided by surface heating or by lifting along a frontal surface gives rise to saturation and to the production of absolute instability in which the particles rise of their own accord. (See page 55.) The vertical currents produced in this way can be very strong, and the contrast between the vertical velocities of currents within the cloud and those in the surrounding air may be very marked. Turbulence in flying from the unsaturated air near a cumulus cloud into the cloud, and that within the cloud itself may be very violent. Generally the small fair weather *cumulus humilis* shows but slight or moderate turbulence, but the well developed cumulus or cumulonimbus may exhibit very strong turbulence. In general turbulence is slightest just above the base of a cumulus cloud or in its highest summits. (Page 240.) If it is impossible to fly above or around a cumulus cloud, therefore, it should be entered near its base and at a slow speed. If conditions are favorable for the formation of hail, as indicated by a well developed *mantle* about the summit of a towering cumulonimbus cloud (see page 288), the cloud should be avoided under all circumstances. Strong cumulus convection, even if apparently unaccompanied by hail, should be avoided at all times since the turbulence may be extremely violent.

Mechanical Turbulence—This type of rough air is the result of eddies produced in the general air flow along a discontinuity surface. It may result from the wind blowing across rough topography or from eddy disturbances along a frontal surface. In either case its effects are generally confined to the immediate vicinity of the discontinuity surface.

Mountainous regions may affect the smooth air flow of the lower 2000–3000 feet of the atmosphere and give rise to irregular bumpiness, generally only of moderate intensity. This occurs when

fairly strong wind velocities are found in the lower several thousand feet, and when the wind blows more or less normal to the trend of the mountains. This type of turbulence is common in the western United States and is frequently observed on a minor scale along the eastern slopes of the Appalachian Mountains on approaching or leaving the coastal plain. This type of turbulence may be avoided by flying a sufficient distance above the terrain to avoid the surface eddies. In the case of strong foehn winds the rough air may extend to a considerable distance above the surface, but this condition is generally easily foreseen and avoided.

Marked turbulence is frequently observed along frontal surfaces. Here the zone of rough air is generally several thousand feet in thickness. Since the frontal surfaces have a very gentle slope in most cases, the horizontal distance which must be traversed in passing through the front may amount to several miles, however. In passing through such turbulent zones it is generally desirable to climb or descend fairly rapidly in order to traverse the rough air as rapidly as possible. In mild fronts, especially those which are not accompanied by heavy cloud systems, the turbulence may be limited to a single bump as the lift on the airplane changes in passing from one air mass to the other. In active fronts the mechanical turbulence due to eddies along them is usually supplemented by thermal turbulence within associated cumulus clouds. In such cases the turbulence may become very severe as in the well developed cold front or *line squall*. Such areas of strong turbulence should be avoided since the vertical velocities and accelerations occasionally reach destructive values.

Passenger Comfort—In commercial aviation the choice of smooth flying levels is of very great importance, since *passenger comfort is directly reflected in passenger travel.* In nearly all cases it is much more important to choose *smooth* flying levels than to choose *fast* flying levels, since the loss of several minutes in a flight is infinitely less important than the acute discomfort of the passengers. Several general rules may be formulated to determine smooth levels from the considerations mentioned in the preceding paragraphs:

1. The air is generally more smooth at night than during the day, since convective activity is less at that time.
2. Turbulence is less pronounced over level country than over

mountainous regions, since disturbances of the smooth flow of the wind are less in the former case.

3. In the case of stratified air, the smoothest levels are generally found just above inversions. Aerographic soundings are valuable in determining such levels.
4. In the case of ordinary convection due to surface heating, the air will generally be smoother several thousand feet above the ground than near the surface unless condensation has occurred with the production of cumulus clouds. In this case the smoothest level will generally be found just below the base of the cloud layer.
5. In flying through regions of cumulus activity the air within the clouds will invariably be considerably more turbulent than the surrounding air since the lapse rate for the saturated state is considerably less steep than for the dry state. It is thus desirable to fly on top or around cumulus clouds whenever possible.
6. In the case of cold fronts, the maximum turbulence is generally experienced near the surface of the ground. In cases where the cold front is displacing markedly unstable air, however, the turbulence may be considerable at all levels below 15,000–25,000 feet.
7. In warm fronts the maximum turbulence is generally found immediately above the warm front surface, and in most cases comparatively smooth air will be encountered below the front.

Most of the above conditions which tend to produce rough air can be anticipated from a study of the synoptic weather chart, together with aerographic soundings for the region. It will be found frequently that by altering the flight path only 1000–2000 feet much smoother air may be found, and although the winds encountered may be more unfavorable at this new altitude, the conditions of greater comfort far outweigh the slight loss in time involved. The pilot may also find that by altering his course a few miles he may be enabled to avoid isolated cumulus clouds or thunderstorms even though the tops of the clouds may be too high to be flown over. Similarly, in mountainous regions it will frequently be found that the mechanical turbulence induced by rough topography diminishes sharply at an elevation of 2000–4000 feet above

the surface and by flying at this altitude relatively smooth air may be encountered when it is rough near the surface.

ICING

The problem of ice formation within the atmosphere is of great practical importance to aviation, since one of the principal hazards to flying in the cold seasons of the year in the higher latitudes is the formation of ice on aircraft. Certain factors governing the formation of ice are fairly well understood and explain most of the observed phenomena in connection with the aircraft icing. Difficulties encountered in accurately forecasting icing conditions arise from the great variability of these factors within relatively small areas.

Aircraft icing is of three general types:

1. *Frost* is a very light crystalline deposit which never develops in large quantities and which is generally experienced only as a thin film over the windows and on the leading edges of the wings of the airplane.
2. *Rime* is a white, opaque, granular structure consisting of very small ice particles which have little cohesion. It tends to form as sharp deposits on the leading edges of the wings and tail surfaces of airplanes and does not greatly alter the form of the airfoils. Since it is but slightly cohesive it is readily shaken off through vibration. Rarely does rime icing constitute a source of danger to an airplane.
3. *Clear Ice* is smooth and glasslike in appearance although it may be somewhat rough when mixed with snow or sleet. It is very cohesive and is only lost with great difficulty. It tends to form on leading edges, but instead of building out in a sharp point, as does rime, it tends to form a mushroom shape which greatly changes the shape of the airfoil. This results in a distinct loss in lift and in an increase in drag, as well as adding to the weight of the airplane. This type of icing may become very dangerous at times.

Since frost and rime are of very little importance to aviation, little will be said concerning their mode of formation. In general they are formed when the water droplets present in the atmosphere are very small. Rime is also more commonly observed when the air temperatures are considerably below the freezing point.

Formation of Ice—Ice may form in either one of two general conditions. In the first case the temperature of the surfaces of the airplane may be lowered well below the freezing point as a result of flying through cold air, and then upon encountering water droplets, generally in the form of rain, which have a temperature above freezing, ice may form upon contact of this water with the cold surfaces of the airplane. This type of icing is frequently encountered in flying through cold air just below a warm front from which rain is actively falling. Icing under these conditions may be quite severe, since if the plane remains within the cold stratum of air its surfaces will remain below the freezing point while a continuous supply of water is available for ice formation.

The second manner in which ice is commonly formed on aircraft results from the presence of *subcooled* water in the atmosphere. Droplets of water which have been cooled below the freezing point without congealing freeze very rapidly upon impact, and thus an airplane flying through a portion of the atmosphere containing subcooled water, may undergo very rapid icing. The manner in which subcooled water droplets can be formed is rather important since their presence and the degree of subcooling affects considerably the icing conditions.

It has been observed frequently in the laboratory that under certain conditions water may be cooled far below its normal freezing point before solidification occurs. This may be accomplished by freeing the water from all impurities and maintaining it in complete rest. A slight agitation of the subcooled water will result in its immediate freezing, also the addition of a minute ice crystal will cause rapid freezing of the entire body of subcooled water. Even the presence of salt particles or dust will greatly hasten the freezing. It is well known that the presence of salt will cause a lowering of the freezing point of pure water. This depression of the freezing point by the addition of dissolved salts amounts to 18.7° C. per gram mole per 100 cc. for undissociated salts and to several times this for well ionized salts. Water, whose freezing point has been lowered by the addition of salts is in equilibrium with its surround-

ings, whereas water which is subcooled due to lack of agitation or to lack of nuclei is very unstable and will freeze when disturbed.

In the case of subcooled water in the atmosphere, each of the means of producing freezing mentioned above may be of some importance, and it will be interesting to determine their relative influence in producing subcooled water and in bringing about its freezing. It is known from the results of several investigators that the salts of evaporating sea spray constitute one of the principal sources of nuclei of condensation, so that each water droplet is thus in effect a salt solution. This fact has been mentioned many times to explain the presence of subcooled water in the atmosphere but from a quantitative point of view it fails completely, for the amount of foreign substances actually present in water in the atmosphere is insufficient to produce a depression of the freezing point of more than a fraction of a degree. The lack of sufficient crystalline nuclei to effect freezing of the water droplets is undoubtedly a partial explanation of the presence of subcooled water, but it is insufficient by itself.

The fact probably is, as shown by a large mass of evidence, that subcooled water in the atmosphere results from ordinary cooling of the water droplets well below the freezing point, and that the *remaining* in this subcooled condition depends upon two factors. The first of these is the lack of sufficient agitation of the drops, and the second is the lack of sufficient *time* for the freezing process to occur. Although laboratory data are insufficient to throw any definite light on the subject, it is fairly well established from a study of icing conditions in the atmosphere with relation to the type of air in which they may occur, that subcooled water is only rarely found in cases where it has remained below the freezing point in excess of four or five days. The equilibrium between liquid water and solid ice thus appears to include the two factors of *time* and *agitation.* Both of these are very important from the standpoint of aircraft icing, since the agitation necessary to effect freezing of the subcooled droplets is provided by the impact of the surfaces of the airplane, and the past history of the air involved controls to a very great degree the amount of subcooled water that may be expected. If the water droplets have remained at a temperature below freezing, beyond the equilibrium period of four or five days, it is unlikely that liquid water will be present at temperatures below the freezing point.

The presence of liquid water at temperatures well below the freezing point has been noted by many investigators. Köhler has reported cloud particles in liquid form at temperatures of $-18°$ F. and Wegener observed water droplets in Greenland at a temperature of $-30°$ F.

McNeal has investigated the question of aircraft icing from the theoretical standpoint and has found that the freezing process of water droplets depends greatly on the degree of subcooling of the water. He shows that the percentage of the droplet frozen on impact depends directly on the degree of subcooling. Thus 10% of a droplet having an initial temperature of 17.6° F. will be frozen on impact while only 5% of a droplet having an initial temperature of 24.8° F. will be frozen on impact. McNeal points out, further, that the *evaporation* of the unfrozen portion of the droplet causes the freezing of a large proportion of the remainder. Thus he shows that of a droplet having an initial temperature of 17.6° F., 10% of it will be frozen on impact, and 78% will be frozen by evaporation, so that of the original droplet 88% will be frozen by the two processes. He also shows that the greater the subcooling, the larger the percentage of each droplet that will be frozen on impact and also the more rapid the evaporation. In arriving at the above figures McNeal assumed a relative humidity of 100% in the cloud air, but he also points out that with lower relative humidities ice formation actually will be aided owing to the more rapid evaporation under such conditions. He gives examples of clear ice formation with relative humidities ranging from 51% to 100% with the average being about 90%. Due to the effect of evaporation in cooling the water droplets, it is possible for ice formation to occur at temperatures above the freezing point, with a temperature of about 35.6° F. being about the upper limit. Ice formation above the freezing point, however, takes place very slowly and any deposits formed under such conditions are unimportant. The formation of ice at temperatures above freezing is also aided by adiabatic cooling of the air in passing over regions of low pressure above certain portions of the airfoil.

It has frequently been stated that rime icing occurs at very low temperatures and that clear icing is only found at temperatures above about 20° F. *This assumption is wholly unfounded and cases of severe clear icing have many times been reported at temperatures well below 0° F.* In fact, the most severe clear icing en-

countered appears to occur under certain conditions at very low temperatures. On the whole, rime icing does occur at somewhat lower temperatures than clear icing. This is as might be expected, since the rapid crystallization which occurs in the case of water droplets which are highly subcooled tends to bring about the instantaneous freezing of an entire water droplet so that there would be no opportunity for the water to spread out after impact. It is this spreading out of the water droplets after impact and while the freezing process is being continued by evaporation that causes the "mushroom" deposits characteristic of clear icing.

McNeal points out that the size of the water droplets is of great importance in determining the type of icing which may occur. Only with large droplets, which can spread out while freezing, as indicated above, can the tenacious mushroom-shaped deposits of clear ice be produced. This corresponds very closely with observational data, which indicate that clear icing is almost invariably found in regions of considerable convective activity, while rime icing is generally encountered in more or less stable air. Thus, the vertical currents necessary to support large water droplets which give rise to clear icing, are closely associated with convection, while the absence of such vertical currents in regions of stable air is associated with very small droplets and rime icing. Pilots frequently observe severe icing, when temperature conditions are appropriate, in flying through cumulus clouds (including stratocumulus and altocumulus types), while only rarely is severe icing reported in stratiform clouds .

From the considerations brought out above it is evident that severe clear icing can occur in either one of two types of synoptic situations:

1. An airplane flying through air below the freezing point, and which encounters water droplets of any temperature, may experience severe icing. This situation generally occurs when flying within the cold wedge of air lying beneath a warm front, from which rain is actively falling. In such a situation the icing conditions will be found to cease abruptly at the warm front surface and any ice deposits acquired below the warm front will rapidly melt in the warm air current above. Frequently, however, icing under such conditions may be sufficiently severe so that it is impossible to climb through the cold air.

2. Icing may be due to subcooled water droplets which may occur at almost any temperature below the freezing point, although they are more common in temperatures above 20° F. This type of situation is restricted to conditions of moderate convective activity with vertical currents sufficient to support comparatively large water droplets, and is generally found only in air masses of relatively recent maritime extraction. It has been pointed out that water can exist in a subcooled state for only a limited amount of time, perhaps four or five days, therefore severe icing in air of continental extraction is very rare. The recognition of unstable conditions within the atmosphere may generally be accomplished by the use of equivalent potential temperature diagrams. *Any unstable layers in an air mass of recent maritime extraction associated with temperatures below the freezing point is sufficient indication for moderate or severe icing conditions.* In the absence of aerographic soundings from which to obtain equivalent potential temperatures aloft, the characteristic air mass properties must be employed and here also the same criteria for icing apply. In this regard it should be noted that the rapid subsidence which usually follows immediately behind fresh outbreaks of polar maritime air, generally limits the zone of icing to the immediate vicinity of the cold front. McNeal points out the following conditions under which convective activity may occur.

(a) Convective action resulting from the heating of surface layers.

(b) Convective overturning or the forcing aloft of warm air at the forward position of the cold front.

(c) Active upglide of warm moist air over a warm front.

(d) The lifting of cold moist air under the warm front due to convergence of streamlines along the warm front.

REFERENCES

UPPER WINDS

1. An Aerological Survey of the United States, Part II. Results of Observations by Means of Pilot Balloons: W. R. Gregg. M. W. R. Suppl. 26, 1926.

This is an extremely valuable summary of upper wind information for the central and eastern part of the United States. The entire region east of the Rocky Mountains is divided into 9 sections. For each of these is given average wind direction and velocity data for the several seasons. The author also presents the information in other ways, including the average percentage frequency of winds from various directions, the average increase in velocity above the surface, etc.

2. Upper-Air Wind Roses and Resultant Winds for the Eastern Section of the United States: Loyd A. Stevens. M. W. R. Suppl. 35, 1933.

This is also a very valuable summary of upper wind information of the eastern United States. Instead of dealing with "average" winds as did the author of the following paper, Stevens deals with resultant winds. These are perhaps more thoroughly representative of conditions that affect flying. This paper treats only the region east of the Ohio River.

3. Winds in the Upper Troposphere and the Lower Stratosphere over the United States: Loyd A. Stevens. Jour. Aero. Sci., v. 4, n. 9, July 1937.

This paper is particularly valuable in summarizing all available information for the United States region regarding winds at the higher levels. The author gives information up to 14,000 meters for selected stations covering the entire United States. The information is presented in the form of wind roses giving percentage frequencies of winds from various directions, and also resultant wind velocities at various elevations at various seasons.

ICING

1. Meteorological Conditions during the Formation of Ice on Aircraft: L. T. Samuels. Nat. Adv. Comm. for Aero. Tech. Notes, n. 439, 1932.

This was one of the first attempts to deal with ice formation on airplanes from the point of view of the synoptic meteorologist.

2. Icing on Aircraft: Edward J. Minser. Bull. Amer. Meteor. Soc. v. 16, n. 5, 1935.

The author deals with the problem from the experience of an airline meteorologist. Well worth while.

3. Ice Formation in the Atmosphere: Don McNeal. Jour. Aero. Sci., v. 4, n. 3, 1937.

This is the outstanding treatment of the subject of icing. It treats the problem from the standpoint of the synoptic meteorologist relating various types of ice to different air masses.

4. Aircraft Icing Zones on the Oakland-Cheyenne Airway: John A. Riley. M. W. R., v. 65, n. 3, March, 1937.

This paper describes in some detail icing conditions along an airway which crosses several high mountain ranges. It is pointed out that most severe icing in this region occurs over the mountains, and during frontal passages. A good, practical paper. Some of the general conclusions with regard to temperatures of clear ice formation do not hold true generally.

5. A Study of Meteorological and Physical Factors Affecting the Formation of Ice on Airplanes: J. K. Lacey. Bull. Amer. Meteor. Soc., v. 21, n. 9, Nov. 1940.

CHAPTER 17

THE WEATHER CHART

INTRODUCTION

The author has had to choose between merely making rather general remarks about map construction, or describing in considerable detail the entire process. He has chosen the latter course. This course has the advantage of describing a certain method of weather chart construction with greater clarity than might otherwise be attainable. It has the distinct disadvantage of almost certainly incurring the displeasure of the meteorologists who use other methods in constructing their charts. It was felt, however, that the setting forth of a definite method of performing all routine operations would be of great aid to the newcomer, and of small hindrance to the experienced forecaster, who would not change his ways anyway!

THE BASE CHART

The base chart used for plotting a weather map is very important and it should be chosen carefully for the exact purpose for which it is to be used. In general, it should be as large as convenient since this makes it possible to employ a relatively large or open scale. If the daily weather chart is to be reproduced by a mechanical process, such as hectographing, the size of the map may be limited by the size of the reproducing machine. As a general rule the scale should not be smaller than 1 inch = 200 miles. Preferably it should be nearer to 1 inch = 125 miles. For oceanic maps, however, the scale may be smaller without affecting the accuracy.

The projection used should be one that does not result in distortion, and which has as nearly as possible the same scale throughout. The Lambert Conformal Conic projection is perhaps the most satisfactory one available for the construction of maps of large areas. The scale is remarkably uniform throughout, and distortion of land

forms is reduced to a minimum. This projection is particularly satisfactory for large longitudinal distances. It is thus particularly well adapted for use in constructing a chart that must include say, the entire Pacific Ocean and portions of the adjoining continents. It is less satisfactory for large north-south distances and generally should not be used for more than 30° of latitude. This projection is particularly simple to construct, since meridians of longitude are straight lines and parallels of latitude are concentric circles. Instructions for constructing a map according to this projection may be found in "Elements of Map Projection," by Deetz and Adams, a publication of the U. S. Department of Commerce, Coast and Geodetic Survey.

In nearly all cases it will be found desirable to depict major topographic features. This may be done in a variety of ways—shading, contouring, hachuring—each one of which has certain advantages. Since the use of either shading or contouring makes it difficult to study the weather information which is entered on the maps, carefully placed hachuring seems to provide the necessary information with the smallest amount of confusion. In general, the meteorologist is more interested in the location of topographic features than in their exact elevation. For this purpose hachuring is as satisfactory as contouring or shading, with the advantages of being more graphic and less confusing, when the weather information is plotted directly over it.

The ink used in printing the map is of some importance. It must be of a color that does not stand out so prominently as to obscure the meteorological information placed on it, yet not be so dull as to be illegible. There are but few advantages to a map printed in several colors and several disadvantages, such as the confusion resulting from a number of colors, and the increased cost. It is generally advisable to choose a single color. Light brown or green colors are widely favored, since they do not conflict with the colors generally used for entering weather information.

Stations should be indicated on the chart by small circles from 1½ to 2 millimeters in diameter. By using circles of this small size, no unnecessary room is taken up on the chart, and yet ample space is provided for indicating the sky condition within the circle, if some care is taken.

The chart should be printed on a good grade of bond paper. Poorer grades will not stand erasing, and do not keep well if the

charts are to be filed for future reference. The surface should take ink well, even after being erased. The entering of weather information is very trying on the eyes and every effort should be made to reduce eye strain. Therefore the paper should be slightly dull in finish. A very light buff or cream color is generally preferable to a pure white.

EXTENT OF AREA TO BE ANALYZED

The size of the region for which forecasts are to be issued determines to a large extent the area which should be included on the weather chart. As a general rule, however, even very local forecasts must be based on weather information covering a rather large area. Generally speaking, rather complete weather data should be available within a radius of 1500 to 2000 miles. This makes it possible to recognize weather disturbances several days before they affect the area for which forecasts are issued. Changes in the structure of the storms as they approach also may be observed. From these considerations, it may be seen that the size of the area for which regular weather analyses should be prepared, is in most cases of at least continental extent. In the vicinity of the seacoast, information on weather conditions in the adjoining ocean is also necessary.

The number of weather reports necessary to construct a satisfactory weather chart varies considerably with the nature of the surface. Over the open ocean, weather reports 300 to 500 miles apart lend themselves to an entirely satisfactory analysis in most cases. Over the surface of the land, especially in mountainous regions, it may be necessary to have weather reports spaced at intervals of 50 to 100 miles, or even less, in order to obtain a satisfactory picture of the weather conditions.

It is desirable to analyze the weather conditions for a comparatively great distance in the direction from which storms generally approach. On the other hand it is comparatively unimportant to analyze conditions for any great distance in the opposite direction. Thus, a meteorologist in the central part of the United States should prepare weather analyses not only for the continental region to the west, but also for the Pacific Ocean within 1000–1500 miles off the Pacific coast, since many weather dis-

turbances affecting the central part of the United States approach from the Pacific Ocean regions. On the other hand, it is very rare that storms originating in the Atlantic Ocean affect the continent farther than a few hundred miles inland, so that a weather analysis extending to the Atlantic coast is generally sufficient in that direction. Since many storms which pass over this area originate in northern Canada, it is desirable that reports be obtained from the entire Canadian area. On the other hand, reports from Mexico are of little interest to the northern Mississippi Valley region, and may generally be omitted.

There should be very little difference between the area covered by a weather chart prepared for strictly local forecasts, and one prepared for forecasts over a considerable area. Thus, whether a forecaster is interested in the entire region of the northern Mississippi Valley, or only in the local weather at Chicago, he will find it necessary to construct a complete weather chart for the United States, Canada and the northeastern Pacific Ocean.

Although it is desirable in the interests of uniformity, that a single base chart be used throughout any weather service, there are also excellent reasons for choosing the most suitable chart for each locality, regardless of uniformity. In the case of a weather service which issues forecasts for a large area, such as the United States, it will generally be found that a single chart to serve properly all portions of the country would have such large dimensions as to be very inconvenient. Forecasters on the Pacific coast would require a map extending to or beyond Honolulu, yet would not be at all interested in the Atlantic Ocean or northeastern Canada. Similarly, forecasters on the Atlantic coast would care little about the Pacific Ocean, but would be very interested in the Atlantic Ocean, at least as far eastward as Greenland. Forecasters in the central part of the country would wish to analyze a portion of the Pacific Ocean lying 1000–1500 miles offshore, and probably none at all of the Atlantic Ocean.

In such a situation, it would obviously be desirable to have three base charts for the proper forecasting of the weather of the United States area. A single chart with an open scale would be too large for convenience. If the scale were to be reduced, or the amount of area included were to be decreased, the resulting compromise would not have the utility of separate, specialized charts for each region.

ENTERING WEATHER INFORMATION

Numerous systems of entering the elements of the weather have been devised, all of which have certain advantages and disadvantages. Generally the meteorologist favors the system to which he is accustomed, although it may not be the most desirable one to an impartial critic. The author has tried most of the systems in common use and has reached certain conclusions with regard to the general features that any system should possess.

First, all signals should be entered in ink, so that subsequent erasing of pencil lines used in preparing the weather analysis may be possible. Second, all signals received at the same time should be entered with a single color of ink. This is a much debated point. Several meteorological services enter the weather signals in 2 or 3 colors of ink. The author has found no marked advantages in the use of several colors of ink to indicate surface weather phenomena, however, when compared to the use of a single color. He has found that the extra time required for this method is a serious obstacle to its use in routine map preparation. Third, the weather information should be grouped around each station in exactly the same manner. Fourth, the various elements should be entered in the order that they are received, since any attempt to enter them in a certain order wastes the time of the person who is entering them. Fifth, one man can enter signals almost as rapidly unaided, as when he has an assistant to call them to him. The latter method of entering signals is wasteful of the assistant's time, which might better be utilized in other duties. Sixth, the wind direction should be shown by an arrow flying with the wind, with the velocity indicated by the number of feathers on the arrow. This method is more pictorial than entering the velocity as a numeral beside the arrow. It falls within the limits of accuracy and representativeness of most wind observations. Seventh, the upper winds should be entered on the same chart as the surface weather information so as to be readily available for use in synoptic analysis.

If the weather information is entered on the base chart in the manner indicated above, it will generally be found that one chart will suffice to depict practically all of the weather conditions of a region. Separate upper air pressure charts, and thermodynamical diagrams are the only exceptions. Both of these can usually be prepared, either just prior to the weather chart, or just after its

completion. They both serve as an aid in the synoptic chart analysis, and should, if possible, be constructed before it.

In certain types of forecasting, where speed is especially essential it may be found helpful to prepare two or more charts, each one of which shows certain of the weather elements. With this plan, several persons may be engaged in working on the different charts at the same time. It will thus be possible to prepare simultaneously, charts of the surface pressure tendency, pressure at upper levels in the atmosphere, etc. When kinematical analyses of the weather conditions are to be made, it will be found very convenient to prepare a separate chart for this portion of the forecasting. Such a chart, portraying the future positions of fronts and pressure systems is of the greatest practical value to the forecaster.

In most cases it will be found that forecasting services employing air mass and frontal technique make it a practice to spend a relatively long time in studying the synoptic situation and in projecting it into the future. This makes it possible for such services to use all of the tools which are available to the forecaster, including upper air pressure maps, kinematical analyses and aerographic soundings, in addition to the standard surface weather chart. This additional time thus spent in a careful study of the synoptic situation, using all available information and technique, is generally found to be very well spent. It results in a delay in the issuance of forecasts, but also almost invariably results in a marked improvement in them.

The details of the system of entering data indicated below have been given the test of extensive use in aeronautical meteorology and in its instruction. Variations in this method are generally unimportant if they fall within the general limitations outlined above.

Upper Wind Data—The upper winds are generally entered an hour or so before the surface information, and should be entered with somewhat longer arrows than the surface winds and in different colors. Each feather on the arrow should represent 10 miles per hour, or 5 meters per second. Half feathers may be used for half units. By using green for the average winds from the surface up to 5000 feet above sea level, red for winds from 5000 to 9000 feet, and blue for winds from 9000 to 13,000 feet above sea level, a very complete picture of the velocity field is immediately available for use in making a complete synoptic analysis. When rapid

changes occur in these levels, two arrows may be drawn for any color, labeling each one with the altitude it represents. Two arrows only, red and blue, representing winds below and above 8000 feet respectively, may also be used, with somewhat less effectiveness.

Surface Data—The amount of information available for use in preparing the weather chart varies considerably. Pressure, temperature, wind direction and velocity, and the state of weather are available in practically all reports. Dew point, barometric tendency and amount of precipitation are also given in most reports. Ceiling and visibility are included in all airways observations. The type of clouds and their direction of movement are given in reports from many stations. Ships at sea in many cases transmit information covering sea swells, sea water temperature, motion of the ship. All of this information is useful in preparing a complete weather analysis and should be entered for every station for which it is available.

The pressure is entered to the right of the station circle, using three figures, the last three of the actual pressure report. Since the pressure is usually reported in millibars and tenths, this manner of entering the report eliminates the "9" or "10." This never results in confusion, however, since the actual range of sea level atmospheric pressures on a single chart rarely, if ever, amounts to 100 millibars. ○112

The temperature is entered above the pressure. The dew point is then placed to the right of the temperature, separated from it by a slant line. The relative humidity is not entered since it is not conservative and is of very little use in analyzing the map. ○12/8

The sky condition is indicated by shading within the station circle.

Unshaded circle—*Clear sky* (less than one tenth of the sky covered by clouds). ○

Half shaded circle—*Scattered clouds* or *partly cloudy* (one tenth to five tenths of the sky covered by clouds). ◐

Three quarter shaded circle—*Broken clouds* (six tenths to nine tenths of the sky covered by clouds). ◕

Completely shaded circle—*Overcast* (over nine tenths of the sky covered by clouds). ●

If the actual percentage of the sky covered by clouds is known, the circle may be shaded accordingly. (Actually the

changes in cloudiness are frequently so rapid that undue care need not be taken in indicating the exact amount of cloudiness.)

The wind direction is given by an arrow drawn away from the circle toward the direction from which the wind is blowing. *The velocity* is indicated by feathers on the wind arrow to represent the force according to the Beaufort Scale, each whole feather representing 2 units, and each half arrow representing a single unit. The feathers may be placed on one or both sides of the arrow, although the force is more clearly shown if they are placed all on one side. When the wind is calm a small "c" is written above the circle.

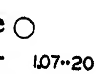
The state of weather is entered to the left of the station circle using international symbols. If the weather condition indicates an overcast sky, the circle should be shaded.

The *visibility* in miles is entered above the *state of weather*, and the *ceiling* in hundreds of feet is entered below it. If no *state of weather* symbol is entered, the *ceiling* will be entered directly beneath the *visibility*.

The barometric tendency is entered below the pressure to the lower right of the station. Both the net change in pressure during the past 3 hours and the tendency characteristic are entered if available. A plus or minus sign need not be used if care is taken to enter the tendency at the top of the symbol for rising pressures, and at the bottom for falling pressures. The *tendency* is usually entered in tenths of millibars.

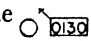
Clouds are entered both above and below the station circle, using international symbols. *Intermediate clouds* are entered immediately above the station circle; their direction of movement is indicated by an arrow at the end of the cloud symbol. *High clouds* are entered immediately above the *intermediate clouds*; their direction of movement is indicated by an arrow at the end of the cloud symbol when no *intermediate clouds* are present. *Low clouds* are entered immediately beneath the station circle; their direction of movement will be indicated by an arrow at the end of the cloud symbol when no other clouds are present.

Amount of precipitation is entered directly beneath the *barometric tendency* in inches and hundredths. A *trace* of precipitation is indicated by underlining the figures 00. *Character*



of precipitation or thunderstorm is entered immediately to the right of the *amount of precipitation*, using international symbols. The *time* of beginning or end of precipitation is indicated to the right of the *character of precipitation.*

Miscellaneous Symbols—A number of other weather elements are occasionally reported, and may be entered on the map if desired. Since the entering of these rarely used symbols should be done in a uniform manner a standard means of indicating them will be pointed out here. *Wind shifts* at individual stations are generally reported by giving the time of the wind shift, and the direction that the wind was blowing prior to its occurrence. This may be indicated by writing the time of the wind shift (using the same time zone for the entire weather chart) just to the right of barometric pressure. A small box should be drawn around the time and a small arrow be attached to one corner showing the direction the wind was blowing prior to the shift. *Maximum wind velocities* may be shown by writing them just above the station circle and drawing a circle around the figures. If the wind direction at the time of a maximum velocity is known this may also be indicated by the use of a small arrow attached to the circle.



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1. Weather Code: U. S. Weather Bureau. Washington. 1939.

This contains the complete numeral (International) code used by the U. S. Weather Bureau in transmitting the data obtained by its stations.

2. International Code for Radio Weather Reports from Ships: U. S. Weather Bureau.

This contains the complete International Weather Code as it applies to ship reports.

3. Decode for use with the International Code for Wireless Weather Messages from Ships: H. M. Stationery Office. London. 1934.

4. Elements of Map Projection with Applications to Map and Chart Construction: C. H. Deetz and O. S. Adams. U. S. Dept. of Comm. Spec. Publ. n. 68, 1934.

This is an authoritative and complete description of all of the commonly used map projections, and contains tables for the construction of many of them.

5. Preparation and Use of Weather Maps at Sea: U. S. Weather Bureau. 1935.

This is an elementary leaflet describing the use of the weather chart by mariners.

6. Distribution of Weather Information, Forecasts, and Warnings by Radio: U. S. Weather Bureau. Radio Circular No. 1.

This lists the various radio services rendered by the governmental agencies of the United States in disseminating weather information. Those interested in obtaining weather signals by radio for use in constructing their own weather maps should obtain a copy of this bulletin.

7. Wireless Weather Messages: H. M. Stationery Office. London.

This bulletin presents the same information for Great Britain that is contained in (6) above for the United States.

CHAPTER 18

WEATHER CHART ANALYSIS

INTRODUCTION

In order to attain real proficiency in the construction of a synoptic weather chart, the student must have a wide knowledge of meteorological theory, a broad background of experience, and a considerable amount of personal instruction by able analysists. Without these prerequisites, the student can never hope to become truly expert in interpreting the weather chart. The present chapter, therefore, is intended only as an introduction to the fascinating field of map construction.

The starting point for anyone who is not familiar with modern weather charts, is the careful study of a number of correctly analyzed weather maps. These should be discussed, if possible, with someone who thoroughly understands them. If possible, the student should read some of the more important of the recent publications, which contain weather map series (see list at end of chapter). Several institutions in the United States maintain complete files of weather charts analyzed according to air mass principles. Their careful study is invaluable.

The student will ordinarily find that he can understand charts of winter weather situations much more readily than summer situations. The contrasts in air mass properties are much greater during the cold season. The pressure systems are deeper and sharper. Fronts are well defined. For these reasons, the beginner in air mass methods should attempt his first weather analyses on charts of the winter months.

PRESSURE CENTERS

The first step in any analysis is to determine the positions of pressure centers and troughs. This information is essential for locating air mass boundaries. If necessary the meteorologist may

sketch the isobars lightly in order to indicate the location of the pressure centers. This practice is not recommended as a general practice, after some ability in map analysis is attained however, as it retards the analysis needlessly. Rather, the analyst should train himself to note mentally the general configuration of the pressure field. He may then proceed at once to locating the air mass boundaries.

FRONTS

Since continuity in the synoptic situation is important in analyses made according to air mass principles, it is very desirable for the forecaster to have in front of him, the preceding weather chart in order that the weather analysis may be carried forward smoothly. It is also desirable to have at hand the weather chart prepared twenty-four hours prior to the one being analyzed, so that the twenty-four hour changes in the various elements may be observed. With these at hand, the meteorologist first attempts to locate the various fronts which were present on the preceding map, in the localities where they may be expected to appear from a normal movement.

The principal lines of cyclonic wind shift should be sketched lightly as a first step in the location of fronts. These lines should then be examined to determine if they represent discontinuities in one or more of the atmospheric elements. The temperature may undergo a sharp drop as the line is crossed. The humidity may fall markedly. An area of low clouds may cease. A region of thunderstorm activity may end abruptly. Any of these changes in connection with a cyclonic wind shift and a pressure trough, is almost unmistakable evidence of an air mass boundary or *front*. In figure 107 some of the various types of discontinuities that may be expected are shown. All of them are associated with a wind shift and a pressure trough. The degree of wind shift and the depth of the pressure trough may vary greatly, but they are always present to some extent.

The pressure tendencies in the vicinity of the suspected front must now be examined carefully. If a front exists, a marked discontinuity in the tendencies bordering it will almost invariably be observed along some portions of it. If the front is moving rapidly, the discontinuity will be very large (figure 108-a). If it is moving

slowly, the contrast will be small (figure 108-b). Only when the front is stationary, will the tendency contrast disappear (figure 108-c).

The analyst should by this time be able to decide whether the fronts that he has located are of the *warm front or cold front* type.

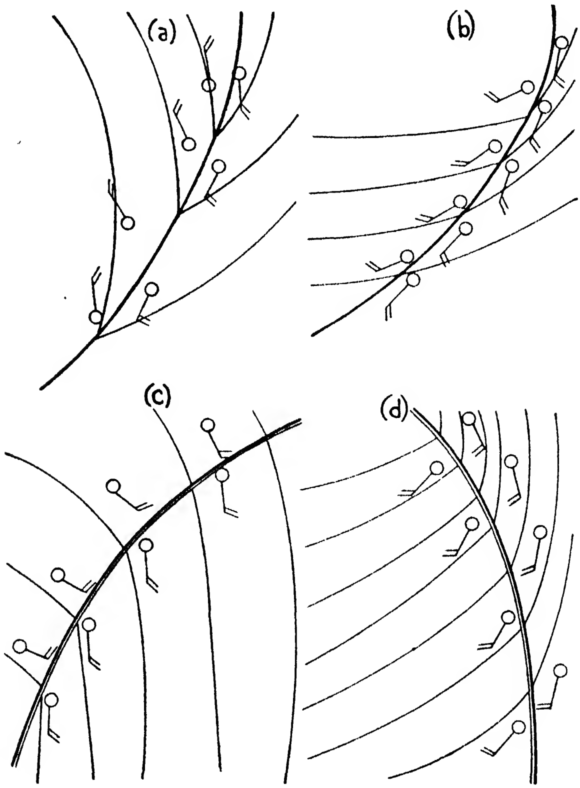

FIGURE 107.—SOME COMMON TYPES OF WIND AND PRESSURE DISTRIBUTIONS NEAR FRONTS

(*a*) Short cold front; (*b*) poorly defined cold front or occlusion; (*c*) poorly defined warm front; (*d*) sharp warm front.

An advancing mass of air that is denser than the air it is displacing will be classed as a *cold mass* and the front as a *cold front*. If the advancing mass is warmer than the air it is displacing, the boundary will be considered a *warm front*. Occasionally there may be some question as to whether a particular front, or a portion of

it, is of a warm front or cold front character. This question often arises when a cold front is turning back as a warm front. In such cases the front, or a portion of it, may be of indeterminate character. It is much more common for a cold front to turn back as a warm front, as indicated in figure 109, than for a warm front to turn back as a cold front. This is simply an expression of the

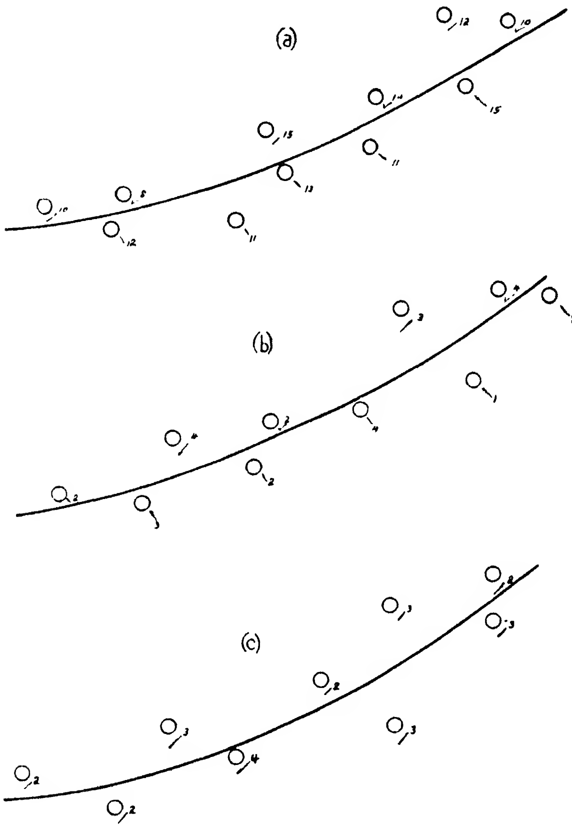

FIGURE 108.—VARIOUS DEGREES OF BAROMETRIC TENDENCY DISCONTINUITY

rule that waves are much more apt to form along a cold than along a warm front.

Since the elements of weather obtained from surface weather reports are subject to many disturbing factors, some of which are of very local effect, much experience is required before the meteor-

ologist can complete an analysis with confidence. This is especially true in mountainous regions where both the wind and the temperature are greatly influenced by topographical effects. In such cases the conflicting testimony of adjacent stations must be weighed carefully in the light of experience. Humidity, when expressed as a conservative property, such as specific humidity or dew point, may be valuable in such cases, although this too may be influenced to a considerable degree by the location of the station. Perhaps the barometric tendency is the most reliable element to be used in mountainous regions in locating fronts. It is certainly more reliable then barometric pressures which have been reduced to sea level. These are particularly subject to large errors during storm conditions, when the temperatures deviate considerably from normal. In stationary fronts, of course, the tendency differential across fronts becomes zero. Their location then becomes very difficult.

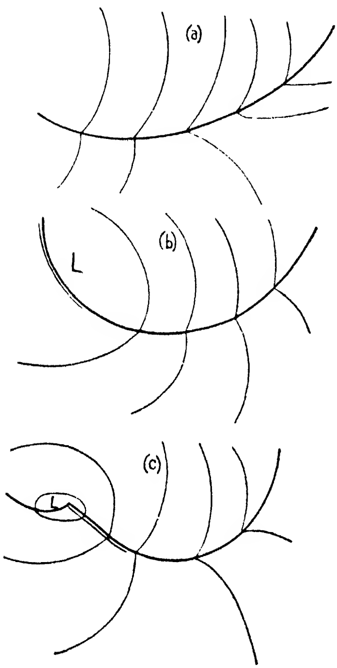

FIGURE 109.—FORMATION OF WAVE ALONG COLD FRONT

Note turning back of left-hand portion of cold front to form a warm front in (*b*) and formation of secondary low pressure center in (*c*).

Upper fronts can in many cases be located by means of surface data alone. Some of the criteria to be used in this connection are pointed out in chapter 9. They should be given careful consideration, since in many cases upper air information may be entirely lacking or at least insufficient to be of much use. In the locating of upper fronts, the plotting of the upper winds on the same chart as the surface data is of great practical value. Discontinuities in the upper streamline field may then be readily correlated with surface phenomena.

WAVES

The pressure tendencies should now be carefully examined in connection with the fronts already located. The tendencies have already been employed in actually locating the fronts, but they

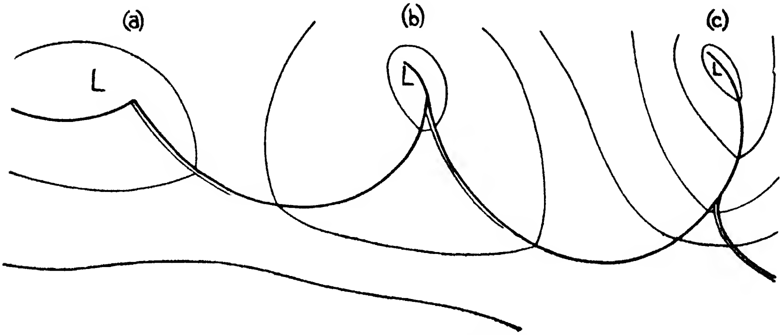

FIGURE 110.—WAVES ALONG A FRONT

(*a*) Stable wave; (*b*) partially occluded wave; (*c*) considerably occluded wave. Note the regular spacing of waves. This feature of constant wave length is especially marked over the ocean where frictional retardation is slight.

should now be used to locate the position of incipient waves. Waves that are well developed will appear with certain characteristic configurations in the frontal system (figure 110). These forms are generally easily recognized. In the earlier stages of their development, however, they may be manifest only by slight irregularities in the frontal surface. The tendencies are of great value in determining whether such irregularities are accidental or whether they represent the first stages in the development of a new wave cyclone. In every case where a new cyclone is in the process of formation, the crest of the incipient wave is marked by a region of generally falling tendencies—an *isallobaric low* (figure 111).

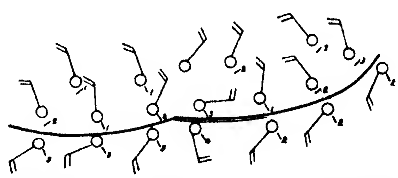

FIGURE 111.—FORMATION OF NEW WAVE ALONG COLD FRONT

Note the tendency toward cyclonic circulation which is developing near the crest of the wave. Also, note the region of falling tendencies near the wave crest, while the rest of the front shows the normal tendencies accompanying a slowly moving front.

When a slight sinuousity along a front is associated with a region of falling tendencies, and if further there are indications of

cyclonic circulation, the analyst is usually well justified in indicating a wave (figure 111). The early recognition of such disturbances is of the utmost importance. Many times these waves travel very rapidly along the frontal surface once they have formed, causing an important precipitation area. In other cases they may move slowly, or even retrograde. Occasionally they may travel a long distance as stable waves. Again they may commence occlusion soon after their formation. Whatever their ultimate history, however, the most important event in their life history is their birth, and the analyst should make every effort to detect it at once.

A useful aid to the detection of waves in their early stages is a knowledge of their average wave length. Waves along the *polar front* in the north Pacific Ocean region thus occur with remarkably constant periodicity (figure 110). Throughout a single season the wave length and velocity of propagation vary but little, at least over the open ocean. As the waves approach the continent, however, they are often retarded somewhat by rough topography along the west coast of North America. It is evident that a study of the average wave length is of great value here in locating newly formed wave cyclones. With the paucity of weather reports and the general absence of pressure tendencies such disturbances often are very difficult to discover.

CHECKING THE ANALYSIS

After the various fronts with their associated waves have been sketched lightly on the weather chart, the analyst should carefully check over his complete analysis. He should make certain that the fronts lie in trough lines, that each wave is in its proper place with relation to the detailed circulation and the isallobaric field. He should be sure that the degree of occlusion of all waves is correctly depicted. There is a tendency on the part of many analysts to indicate the amount of occlusion in a very careless manner. Actually the occlusion usually ends at a low pressure center (figure 112-a). Occasionally, however, it may continue through the center of lowest pressure, particularly when the trough line is sharply marked (figure 112-b). With stable waves, no occlusion at all exists. Care should therefore be taken to indicate the correct

amount of occlusion and whether it is of the *cold front* or the *warm front* type. If of the warm front type the *upper cold front* should be indicated, since it is usually of more synoptic importance than the surface position of the warm front occlusion. The tendency field in advance of an occlusion should always be examined for signs of the upper front (page 206).

If the distance between adjacent waves appears to be unduly great for the season of the year, the region between them should be examined carefully for signs of another wave. If a front is practically stationary, or is being retarded, it should also be examined for possible incipient waves. As has been pointed out, these may develop very rapidly and greatly affect the development of the synoptic situation.

Cold bodies of air should be examined carefully for *important* secondary fronts. Often the inexperienced analyst tends to err on the side of overanalysis in this regard. In an active outbreak of cold polar air during the winter, an almost infinite number of so-called "secondaries" may be discovered. Of these only a few are actually of synoptic importance. These few important ones must not be overlooked, however.

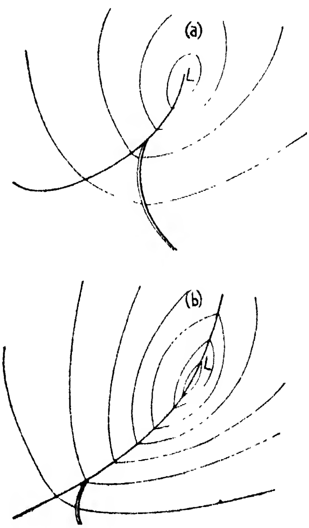

FIGURE 112.—RELATION OF LOW CENTER TO CREST OF OCCLUDED WAVE

(*a*) Usual case, with occlusion ending in center of lowest pressure. (*b*) Rare case, with occlusion continuing through low pressure center. Often a wave develops at the LOW in this case. This case usually occurs with very sharp pressure troughs.

ISOBARS

After having sketched all the elements of the frontal analysis, the isobars may be constructed. They should be drawn with a soft pencil, using a fairly firm touch. They should never be drawn in ink until the map is entirely completed, for even the most accomplished analyst will find it necessary to use his eraser almost more

than his pencil in obtaining a completely smooth pressure field. The first few isobars should be sketched in very lightly for they will almost invariably be erased several times before they are located correctly. In general an isobar that is more or less intermediate in the general pressure field should be drawn first. This isobar will generally appear at many localities over the chart and will serve as a guide to the location of others. The 29.9″, 30.0″ or 30.1″ isobar is generally constructed first for this reason.

All isobars show an *inflection* at frontal surfaces. This is to be expected since a front represents a discontinuity in air mass properties, and it would thus be entirely accidental if an isobar should have the same direction in both air masses. This inflection of the isobars at fronts is one of the principal reasons for constructing the frontal system before drawing the isobars. Figure 113 illustrates some of the more common isobaric patterns in the vicinity of various types of fronts and pressure centers. (See also figures 107, 109, 110, 112, 119, 120.) It will be worth while for the beginner in weather map construction according to air mass principles, to study these patterns. They do not by any means represent all possible cases, but they will serve as a useful guide to many frequently encountered situations.

It may be found necessary at first, when preparing analyses according to air mass principles, for the meteorologist to draw the pressure field before locating the positions of discontinuity surfaces. This is not to be recommended as regular practice, however, since it should be possible for the analyst to ascertain the positions of high and low pressure areas on the weather chart without actually drawing the isobars. Since the pressure field must be drawn to fit the fronts, it is obviously a waste of time to draw the isobars and then modify them to fit the fronts after they have been drawn. The experienced analyst thus locates the position of all fronts on the map before drawing the isobars.

Smoothing the Pressure Field—One of the most important details of the isobaric analysis is the careful *smoothing* of isobars. *Smoothing* is a method of eliminating errors in pressure measurements, during the actual construction of the isobars. If an isobar were to be constructed by joining with straight lines adjacent stations having the same barometric pressure, the result would be an entirely *unsmoothed* isobar. Such an isobar probably would never

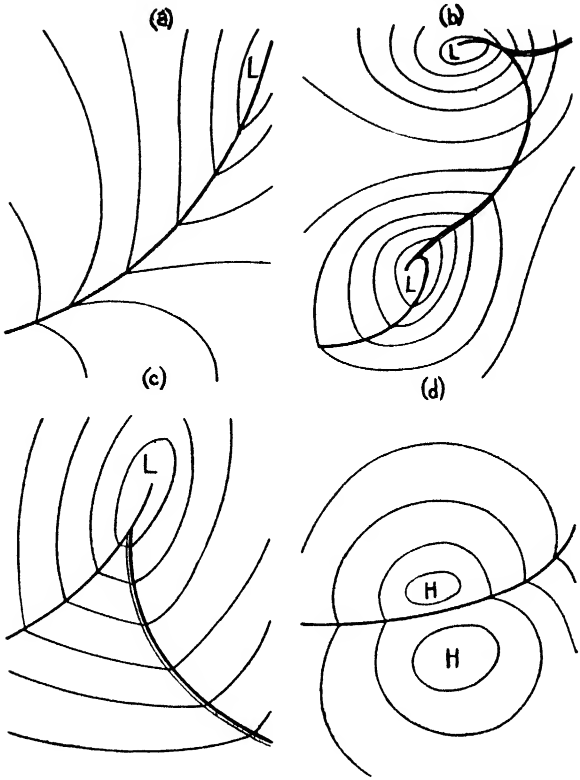

FIGURE 113.—VARIOUS TYPES OF THE MORE COMMON ISOBARIC PATTERNS APPEARING ALONG FRONTS

(*a*) Well defined cold front; (*b*) well developed occluding waves; (*c*) crest of partially occluded wave; (*d*) slowly moving cold front breaking through HIGH.

represent the actual pressure distribution because of several reasons:

First, the individual reports may not have been strictly synchronous;

Second, their reduction to sea level may not have been entirely satisfactory (this is particularly true with high altitude stations);

Third, the observer may have not read the barometer correctly;

Fourth, the station barometers may not have indicated correctly due to lag or other instrumental defects;

Fifth, the reports may have been transmitted incorrectly;

Sixth, the reports may have been encoded or decoded incorrectly, if a code was used in transmitting them.

It is quite evident from the above discussion that an unsmoothed isobaric field will in nearly every case give a false conception of the actual pressure distribution. The analyst must therefore draw his isobars with due regard for the various disturbing elements noted above. In low lying regions, near the seacoast, comparatively little smoothing will be found necessary since the reduction to sea level will be unimportant as a source of error. In mountainous regions, however, a great deal of smoothing may be required, since the errors involved in sea level reduction may be very large.

Several general rules for drawing isobars are given below. All of these may not apply in every case, but they will serve as a basis for the construction of a representative pressure field.

1. Individual isobars should be smooth curves.
2. They are always continuous lines—never discontinuous as isallobars may be.
3. They should not show any random sinuousities unless these are reflected in adjoining isobars.
4. In general, adjoining isobars should have very nearly the same configuration. The more intense the pressure, field, the more valid is this rule.
5. With intense pressure fields, it will be found desirable first to draw every other isobar. After these have been drawn, the intermediate ones may then be constructed. This results in a smoother pressure field than if each isobar were to be drawn in order.
6. Each isobar must show an inflection as it crosses a front.
7. Isobars tend to be crowded when an air mass encounters an obstacle such as a mountain range.
8. Surface winds at sea may be used as a very reliable guide to the direction of the isobars here. Frictional forces deviate the wind relatively slightly from its gradient direction.
9. Surface winds over the land are not especially reliable as indicators of the isobar configuration except over level country. Even here the deviation from the gradient direction is considerably greater than at sea. Where upper wind data are available it will be found desirable to indicate on the weather chart the direction and ve-

locity of the wind at the gradient level. The isobars follow these winds very closely.

10. Remember that an eraser is essential. The more it is used, the more accurate the pressure field will be. For kinematical analysis (chapter 19), a smooth and accurate pressure field is an absolute essential.

ISALLOBARS

When the field of pressure tendencies is to be constructed, it should generally be drawn very lightly with a thin lead, red colored pencil. The isallobars should be drawn after the fronts have been sketched, and either before or after the isobars have been drawn. As with isobars, the isallobars should be smoothed very carefully. This is particularly true immediately to the rear of moving fronts. Here the tendency will invariably be reported too low, and this fact must be given consideration in drawing the isallobars. The isallo-

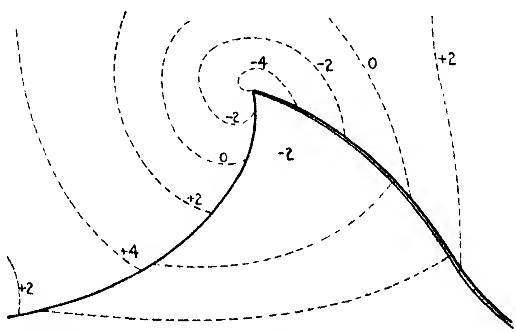

FIGURE 114.—CONSTRUCTION OF ISALLOBARS IN A WARM SECTOR CYCLONE OF TYPE "B"

Note that they are discontinuous at the fronts, both of which are moving. Also note the isallobaric LOW at the wave crest, indicating deepening and approaching occlusion.

bars are discontinuous at moving fronts (figure 114). In general isallobars should be drawn for every .02 inch pressure change, although in flat isallobaric fields they may be constructed for each .01 inch pressure change. The centers of isallobaric "HIGHS" and "LOWS" may be indicated with a red plus sign or blue minus sign respectively. The zero isallobar should always be drawn first, since it is the most important one. Even if the complete tendency field is not drawn, it will often be found desirable to draw the zero isallobar.

COMPLETING THE MAP

After constructing the pressure field the meteorologist needs only to indicate the various types of fronts and the areas of precipitation to complete the weather map. The following colors and patterns are widely used for the various fronts:

Cold front—solid blue line.
Warm front—solid red line.
Occluded front (cold front type)—solid purple line.
Occluded front (warm front type)—red dash and dot line.
Upper cold front—dashed blue line.
Upper warm front—dashed red line.
Upper occlusion—dashed purple line.
Front of doubtful location—dotted line.
Cold front changing to warm front, or vice-versa—alternate red and blue dashes.

All of the above lines should be drawn with either a heavy pencil, or a broad pen, so that the lines produced are about 1/16″ wide. Unless reproduction of the map by some form of hectograph process is contemplated, it will generally be found desirable to indicate the fronts by means of soft colored pencils. Alterations in the analysis, as brought out by further developments, may then be made readily if desired.

Precipitation—In most cases only precipitation which is occurring at the time the map observations are taken, is indicated on the synoptic weather chart. Past precipitation is not synoptic and therefore should not be shown. Several types of precipitation may be distinguished and each of them should be indicated in a slightly different manner.

Continuous Heavy Precipitation, as that which accompanies many warm fronts, may be indicated by solid green shading.

Intermittent Precipitation, such as that which accompanies weak fronts of any type, where the distribution is rather erratic, may be shown by a series of closely or widely spaced green lines, depending on the intensity of the precipitation.

Showers, such as occur in unstable air behind a cold front, may be shown by a series of small green triangles with their bases upward.

Orographical Precipitation, such as may occur along the windward slopes of mountain ranges when moist air blows across them, may be shown by a series of long and short green lines with their spacing indicating the intensity of the precipitation.

Drizzle Conditions, such as occur in tropical maritime air masses which move to high latitudes, may be shown by a series of green drizzle symbols, as indicated by the International Code.

Movements of Pressure Centers—Whether or not a kinematic analysis of the weather chart is carried out, it is desirable at least to estimate future positions of pressure centers occurring on the synoptic chart. This information should be indicated on the map as follows:

High Pressure Centers—A red arrow should emanate from the present center, with the head of the arrow indicating its position 12 or 24 hours hence. The pressure at that time should be indicated at the head.

Low Pressure Centers—A blue arrow may be used exactly as with the high pressure centers.

Movement of Fronts—As with pressure centers it is also desirable to know the future position of fronts. For this purpose, their position 12 or 24 hours hence should be indicated by a dashed brown line. Even if a kinematic analysis is not carried out the future position should at least be estimated.

Regions of frontolysis may be indicated by the use of the symbol *FL,* and regions of frontogenesis by the use of the symbol *FG.*

Air Masses—Local terms should generally be used for weather charts which are employed in routine forecasting. Modifications of air masses may be indicated either by the use of Willett's symbol for transitional air masses, "N", or by using Krick's notation, which indicates the number of days that the air mass has spent over land and water surfaces since leaving its source region (see page 104). Red symbols should be used for air masses possessing relatively high entropy, and blue symbols for those with relatively low entropy.

REFERENCES

1. Isentropic Analysis: C.-G. Rossby and collaborators. Bull. Amer. Meteor. Soc., v. 18, n. 6, 1937.

This is a concise statement of a method of weather chart analysis recently developed at M. I. T. by Rossby and some of his colleagues. It is supplementary to the ordinary weather chart as prepared according to the Norwegian principles, not replacing the latter in any regard. Actually it presents the topography of a certain surface of constant potential temperature, with isolines of specific humidity superimposed on it. The practical value of this method of analysis has been fully demonstrated. It shows very clearly the movement of tongues of moist air and their connection with precipitation. For this reason it makes it more readily possible for the forecaster to decide on the probable movement of rain areas, than by the use of the surface chart alone. Several important features of the circulation of the middle latitudes are explained better by the use of isentropic charts than in any other way. For climatic studies their value will undoubtedly be found to be very great. See the reference at the end of chapter 12 to the article by Namias.

See references at end of chapters 6, 9, 11. Most of these describe synoptic situations of general interest.

CHAPTER 19

FORECASTING

INTRODUCTION

The construction of a weather chart and its analysis are only the first steps in the making of a weather forecast. The actual preparation of the forecast involves the projection of this analysis into the future, and the determination of the changes that will occur in the present weather. These two steps are by all odds the most difficult in the forecasting process, involving as they do the exercise of much personal judgment. Experience plays a more important part here than in any other step of forecasting.

Great improvements have been made recently in the technique of map analysis. The meteorologist can calculate the development and movement of pressure systems with remarkable success. Nevertheless, forecasting has by no means reached the point where it is mechanical. The complexity of weather processes is so great, in fact, that it is doubtful if forecasting will ever be on that basis. Wide experience in practical forecasting serves to bridge to some extent the gap between the present weather and the changes that it will undergo in the future. The experienced forecaster can thus foretell many of these changes largely from his past experience. This does not mean at all that there should be a blind following of empirical ideas, but rather that each weather situation should be studied in the light of former situations. In this regard, the forecaster must consider all of the weather elements that are involved, and not merely the pressure field.

One of the most important phases of forecasting is a careful review of each weather situation, and an attempt to explain all weather phenomena which have occurred. This so-called, "after-casting" is one of the best ways of discovering forecasting errors and in assuring that similar mistakes will not be made in the future. Unless this critical review of all weather situations is undertaken regularly, the forecasting ability of the meteorologist will show but

little progress. A careful study of "unexpected" weather changes will generally disclose that a proper analysis of the weather chart at the time the forecast was prepared would have made them evident. This will prevent in many cases the recurrence of such mistakes.

It is hardly possible to summarize in a single chapter all of the factors which should be considered in practical forecasting. Each chapter in this book contains more or less information which is useful in determining the way in which weather changes occur. The present chapter is concerned mainly with a discussion of means of calculating changes in pressure and frontal systems. A number of practical rules for forecasting such changes are also included.

It has been mentioned before that one of the most difficult problems confronting the forecaster is the projection of the synoptic weather chart into the future. No matter how carefully the chart may be analyzed, the forecaster will have but little success if he cannot accurately picture the weather situation 12–36 hours hence. The original analysis is difficult enough, but its projection into the future may be a problem of the greatest complexity. Not only must the major details of the original analysis be correct, but their future position and intensity must be determined as well.

Although the projection of every detail of the weather chart is a manifest impossibility, yet the forecaster should endeavor to determine the future position of all pressure centers and the more important fronts, together with changes in their intensity. Petterssen's outstanding work on this subject will form the basis of the present chapter, which will describe some of the more recent research concerning the kinematics and dynamics of the pressure field.

REPRESENTATIVE PRESSURE

Since barometric pressure is the only meteorological element which is truly representative of a given locality, it is used by Petterssen in his investigation. Practically all of the material in this chapter is based on atmospheric pressure. This should suffice to emphasize the importance of obtaining barometric observations about whose accuracy there can be no doubt. The more common sources of error in pressure measurements, and methods of smoothing the pressure field to eliminate them are described on pages 356-358. The

isallobaric field is also very important in a kinematic analysis of the pressure field, and it too must be constructed, at least in part, if satisfactory results are to be obtained. Careful smoothing of the isallobars is again essential (page 359).

PRESSURE PROFILES

Several terms are used in this discussion with which the reader should be thoroughly familiar. A *pressure profile* represents the distribution of pressure along a line of a weather chart. It is in all respects similar to a profile of a land surface. Figure 115 shows

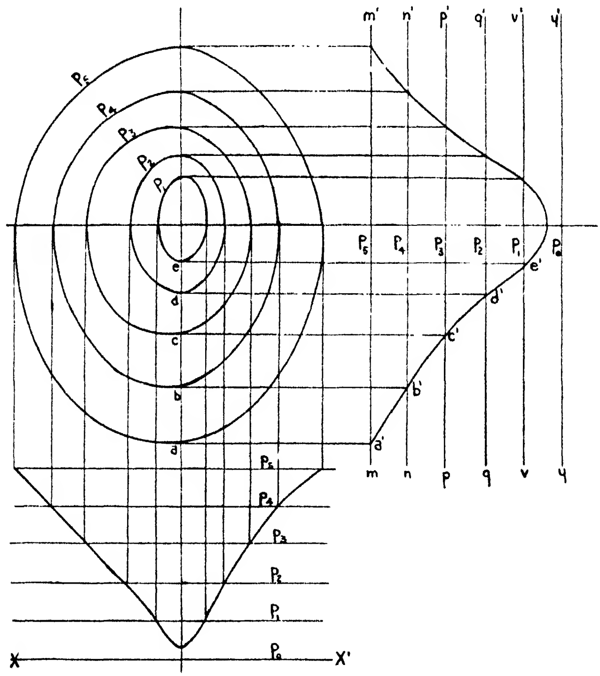

FIGURE 115.—CONSTRUCTION OF PRESSURE PROFILES

a low pressure area as outlined by isobars. Beside it are *pressure profiles* along two perpendicular axes. The profiles are constructed by first drawing the axes along which they are to be constructed (xx and yy). At a convenient distance, or on another sheet of paper draw the construction axes (xx′ and yy′) parallel to the

original ones. Drop perpendiculars from the intersection of the isobars with the original axes, to the construction axes (aa′, bb′, cc′, . . . etc.). Choose convenient units to represent the pressure gradient and draw lines parallel to the construction axes to represent them (mm′, nn′, . . . etc.). Ordinary cross-section paper is convenient for this purpose. The profile may then be drawn through the intersection of the construction lines, as shown in figure 109. If cross-section paper is used for constructing the profiles one edge of a sheet may be placed along one of the axes (xx′ or yy′) and the profile may be drawn immediately.

In order to designate in an abbreviated form, certain properties of the pressure field, Petterssen has developed a useful notation.

p_{100} *means the change in pressure per unit length along the x-axis.*

p_{010} *means the change in pressure per unit length along the y-axis.*

These quantities thus express the *slope of the profile.* They are positive when the profile *ascends* toward the positive end of the x- or y-axis; negative when the profile *descends* toward the positive end.

p_{200} *means the change in slope per unit length along the x-axis.*
p_{020} *means the change in slope per unit length along the y-axis.*

These quantities express the *curvature of the profile.* They are zero when the slope is constant, large when it is considerable. The profile is said to have a *cyclonic curvature* when it is *concave upward,* an *anticyclonic curvature* when it is *concave downward.* The profile in figure 118 is curved anticyclonically at the left and right extremities, and cyclonically near the center.

TENDENCY PROFILES

In the same way that profiles are drawn of the isobaric field, they also may be drawn of the field of barometric tendencies or isallobars. These are called *tendency profiles.* Figure 116 shows a field of isallobars and the corresponding profile along the x-axis.

Petterssen's notation for features of the tendency profile is as follows:

p_{101} *means the change in the tendency per unit length along the x-axis.*

p_{011} *means the change in the tendency per unit length along the y-axis.*

These quantities thus express the *slope of the tendency profile.* Their sign is the same as for the pressure profiles.

p_{201} *means the change in slope per unit length along the x-axis.*
p_{021} *means the change in slope per unit length along the y-axis.*

The quantities express the *curvature* of the tendency profile. The terms *cyclonic* and *anticyclonic* have the same meaning as for pressure profiles. Thus, the left hand portion of figure 116 is curved anticyclonically, and the right hand portion cyclonically.

Generally, the tendency profiles are constructed on the same axes as the pressure profiles. They are then directly comparable. The origin of the axes is usually a pressure center, or at least a trough or ridge line. The x- and y-axes follow the symmetry lines of the pressure system. As a result of this construction, a point of inflection along the pressure profile is located on one axis. This is rarely the case with the tendency profile, however, for only rarely does the isallobaric low coincide with the isobaric low.

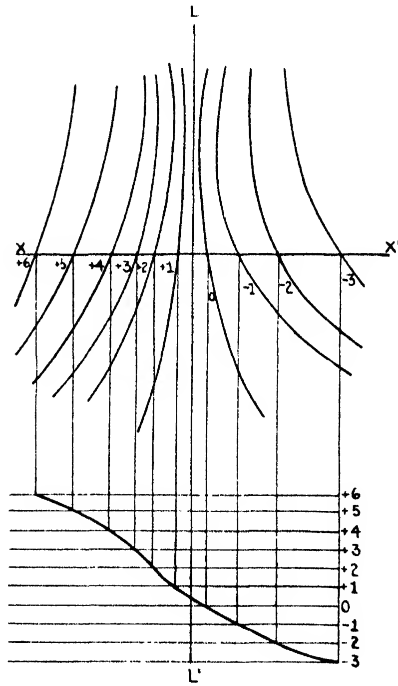

FIGURE 116.—FIELD OF ISALLOBARS AND CORRESPONDING TENDENCY PROFILE

The *ascendant* of pressure is defined as the *increase* per unit length of the pressure in the direction toward which it *increases most rapidly.* It is *perpendicular* to the isobars, points toward a region of *higher* pressure and its magnitude is *inversely proportional* to the distance between the isobars.

The *gradient* is defined as the *decrease* per unit length in the direction where the *decrease* is *most rapid.* It is therefore equal in magnitude to the ascendant, but points in the opposite direction.

The ascendant or gradient of the tendency have precisely the same meanings as for the pressure. The *isallobaric ascendant* thus points toward high values of tendency, and the *gradient* toward low values.

VELOCITY OF ISOBARS

The normal velocity of an isobar may readily be calculated if the distance between unit isobars, and the barometric tendency are known,

(1) $$V_i = - Th$$

where, V_i is the normal velocity of the isobar, T the barometric tendency, and h the distance between unit isobars (figure 117).

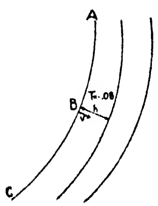

FIGURE 117.—VELOCITY OF ISOBARS

T = local tendency; h = distance between adjacent isobars; V_i = instantaneous velocity = $-Th$.

The direction of movement is to the right when the result is positive and to the left when it is negative. *It is important that the same units be used throughout.* In cases where the tendency is reported in hundredths of an inch, and the pressure field is drawn for each tenth of an inch, care must be taken that this fact is considered.

Example. Calculation of isobar velocity.

Given: Distance between isobars (h) = 50 miles.

Local tendency (T) = -8

Required: Normal velocity of point B on isobar ABC (figure 117).

Solution. Reducing the given quantities to the same units, the distance between unit isobars is 50 miles and the tendency is -0.8. (Both the distance between unit isobars and the tendency refer to *tenths of an inch,* so the same units are used throughout.)

$V_i = -(-0.8 \times 50)$

$V_i = +40$ miles per three hours, to the right.

This formula may be used for determining the *instantaneous* velocity of any portion of an isobar, and also for extrapolating the movement of it for a moderate period of time. The length of time

for which such an extrapolation may be made depends on the acceleration of the isobar, as brought out in a later section. Actually, while the tendency and the pressure gradient may vary widely and rapidly at a given ground station, the velocity of an element of a moving pressure system remains comparatively constant with respect to the *moving system.*

This elementary formula is actually of considerable use in making routine forecasts. It is easily possible to estimate the change in wind velocity accompanying a moving pressure system by calculating the velocities of neighboring isobars and thus determining the variation in pressure gradient. For this purpose the formula below, derived from the Velocity formula, may be used,

$$h = h_o + (V_2 - V_1)t \tag{2}$$

The pressure gradient is equal to $1/h$, so that if the wind velocity is assumed to be proportional to the pressure gradient, a quantitative forecast of the wind velocity may be made with the use of the above formula.

Example. Calculation of change in pressure gradient.
Given: Initial value of distance between adjacent isobars (h_o) = 50 miles. Velocity of one isobar = 40 miles per 3 hour interval. Velocity of other isobar = 45 miles per 3 hour interval. Time = three 3 hour intervals.
Required: Value of h after given time.
Solution: $h = 50 + (40–45)\ 3$
$= 35$ miles (whether the quantity in parenthesis is to be added or subtracted may be determined by inspection).

If the wind initially has a velocity of 25 miles per hour, and if its velocity be assumed to be proportional to the pressure gradient, then the velocity after 9 hours will be:

$$V = 25 \times \frac{50}{35}$$

$$= 36 \text{ miles per hour.}$$

VELOCITY OF ISALLOBARS

An *isallobar* is a curve representing equal values of the *pressure tendency*. It is analogous to an isobar, but it represents pressure tendency instead of pressure. The normal velocity of an isallobar may be calculated if the distance between unit isallobars and the rate of change of the tendency are known,

$$V_t = -\frac{\partial T}{\partial t} H \tag{3}$$

where $\frac{\partial T}{\partial t}$ is the rate of change of the tendency, and H is the distance between unit isallobars. This formula is thus very similar to the one used to determine the velocity of an isobar.

ACCELERATION OF ISOBARS

The preceding formulae give only the instantaneous value of the velocity of an element of an isobar or isallobar. While this will yield results that are satisfactory for short extrapolations, it is necessary to know the acceleration of the element to gain an accurate knowledge of its future movement.

The acceleration of an isobar may be calculated by the use of the following formula, which applies to the x-axis:

$$A_i = \frac{P_{002} + 2V_i P_{101} + V_i P_{200}}{P_{100}} \tag{4}$$

The meanings of all of the above terms except P_{002} have been explained above. P_{002} *is the rate of change of the barometric tendency.* It can be obtained from two successive weather charts. All of the other terms may be obtained from a single chart.

DISPLACEMENT OF ISOBARS

After the acceleration is obtained, the total displacement may be computed from the usual formula for motion involving velocity and acceleration:

$$S_i = V_i t + \tfrac{1}{2} A_i t^2 \tag{5}$$

where S_i is the total displacement of the isobar during the time t.

NUMERICAL DIFFERENTIATION

Petterssen's notation for the various elements of the pressure field is based on actual differentiation. This is hardly practical for routine calculations, so a system of numerical differentiation has been devised that employs data readily obtainable from one or two weather charts. Profiles of pressure and tendency are first constructed. Units are then marked off on the profiles as illustrated in figure 118.

The choice of axes to be used in computing pressure center movements is a very important matter. In fact, the success or failure

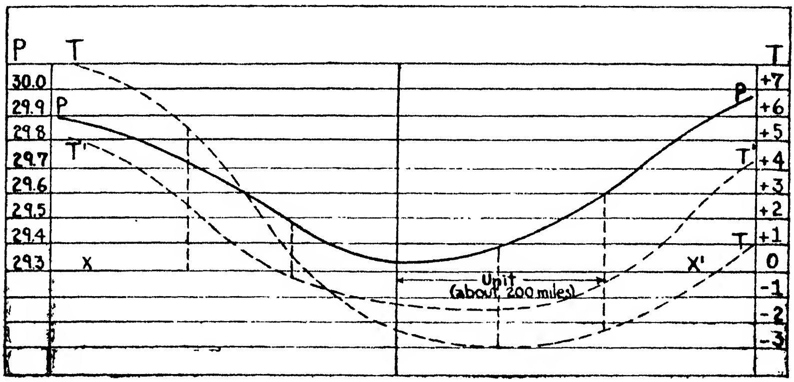

FIGURE 118.—METHOD OF NUMERICAL DIFFERENTIATION APPLIED TO PRESSURE AND TENDENCY PROFILES

of a forecasted movement very often hinges on this choice. In many cases the axes used may simply be the axes of symmetry of the pressure centers. However, when the isallobaric ascendants (or gradients) deviate materially from the symmetry axes, it is necessary to use the former as computation axes. Usually it will be found that there is a major isallobaric axis running from a main anallobar through the center to a katallobar. Nearly perpendicular to this axis is another one running from a smaller anallobar through the center to a smaller katallobar. It is convenient to use the major isallobaric axis as the x-axis, and the minor one as the y-axis. In computing the movement of troughs or wedges only one axis is used, of course. This one should be constructed perpendicular to the trough or wedge at the point selected for computation.

The size of unit is also very important in practical work. The proper size of unit will usually yield good results, whereas a unit that is too short or too long may lead to a wholly erroneous computation. As a general rule, the unit should be rather large for flat pressure gradients and isallobaric fields, and fairly small for intense fields. Other matters must be given careful attention, however, in actual forecasting. First, the unit must always be equal to at least twice the 3-hourly movement of the system. This eliminates at the outset errors due to choosing tendencies too close to a rapidly moving system. Second, the unit must never be so long as to become involved with an adjoining system. If systems are so close together that these two requirements cannot be met simultaneously, the alternate formulae in (1), (2), (5), (6) below may occasionally be used to advantage. With extremely closely spaced systems it may be necessary to consider two or more small systems as a single larger one and perform calculations on the entire system as a whole.

If the region near the center of a system is unusually complicated by fronts, it is often desirable to choose *supplementary axes* parallel to the principal axes through the center of the system. These should be located a short distance on one side or the other of the main axes, so that they are away from frontal influences. Such supplementary axes often allow the analyst to obtain representative tendencies when he cannot obtain satisfactory values in the immediate vicinity of the principal axes due to the confusion of data near them. When supplementary axes are used in this manner to obtain representative tendencies, the pressure data used in the kinematic formulae should be obtained as usual from the principal axes.

When the tendency profiles are not symmetrical about the pressure center, care must be taken to avoid using too long a unit. Thus, in figure 118 the tendency profile TT is asymmetrical about the LOW center. The katallobar is located only a short distance to the right of the center, while the anallobar is located far to the left. The unit as chosen in this case is the *longest* possible one. A longer one than that shown will give an erroneous measure of the tendency profile slope using formula (1) below. If the unit were made any longer in this case, the computed slope of the tendency profile would be too small. Often the longest possible unit in an asymmetrical system is also approximately the *smallest* possible unit, due to the necessity of choosing a unit at least equal to twice the 3-hourly

velocity of the system (using formulae (1) or (2) below to compute the slope of the tendency profile).

In deciding on the actual values of the tendencies to be substituted in the formulae for numerical integration below, considerable care must be taken to choose thoroughly representative values. Values which show little agreement with those at surrounding stations should be eliminated. Especial care must be taken in thunderstorm areas, since the tendencies are usually very unreliable there. An isallobaric field may be constructed if desired, in order to obtain the most representative values of the tendency. Usually, however, a careful study of all tendency values in the immediate vicinity of the point under consideration will lead to the choice of a truly representative tendency. Only the poorest results may be expected if the analyst merely uses blindly whatever value of the tendency happens to appear on the surface map beneath the axis at the point being used as a unit or half-unit of length.

It will be found as a general rule that in using velocity formulae such as (1), page 374, that the results will be relatively unreliable if the denominator is smaller than about 10 millibars, and that results will be entirely unreliable if it is smaller than about 5 millibars. This means practically that the kinematic formulae are not useful when dealing with flat pressure fields.

The equations below give the numerical formulae for calculating all of the various expressions for the pressure and tendency field:

(1) $P_{101} = T^{\frac{1}{2},0} - T^{-\frac{1}{2},0}$ or $\frac{1}{2}(T^{1,0} - T^{-1,0})$

(2) $P_{011} = T^{0,\frac{1}{2}} - T^{0,-\frac{1}{2}}$ or $\frac{1}{2}(T^{0,1} - T^{0,-1})$

(3) $P_{200} = P^{1,0} - 2P^{0,0} + P^{-1,0}$

(4) $P_{020} = P^{0,1} - 2P^{0,0} + P^{0,-1}$

(5) $P_{102} = \Delta T^{\frac{1}{2},0} - \Delta T^{-\frac{1}{2},0}$ or $\frac{1}{2}(\Delta T^{1,0} - \Delta T^{-1,0})$

(6) $P_{012} = \Delta T^{0,\frac{1}{2}} - \Delta T^{0,-\frac{1}{2}}$ or $\frac{1}{2}(\Delta T^{0,1} - \Delta T^{0,-1})$

(7) $P_{201} = T^{1,0} - 2T^{0,0} + T^{-1,0}$

(8) $P_{021} = T^{0,1} - 2T^{0,0} + T^{0,-1}$

The alternative equations to the right are to be used when the unit of length is small. In each case, the first coordinate (as $T^{\frac{1}{2},-}$) refers to the x-axis, and the second (as $T^{-,0}$) refers to the y-axis.

Movements of Troughs or Wedges—A *trough* or a *wedge* may be defined as an elongated low or high pressure area. It is not necessarily associated with closed isobars. The curvature of the isobars is a maximum along the trough or wedge line. Since troughs and wedges may be treated in the same manner, and by the use of the same formulas, they will be discussed under the heading of troughs only.

The formula for the *velocity* of movement of a trough may be given as:

$$(1) \qquad V_L = -\frac{P_{101}}{P_{200}}$$

where V_L gives the *normal* velocity of the center line of the trough. This equation neglects all terms of higher order than the first, but it is satisfactory for most practical purposes. When the trough line rotates during its movement, the above equation will not be strictly valid, although the actual discrepancies are only important when the trough is poorly defined.

The *acceleration* of a trough line is given by formula:

$$(2) \qquad A_L = -\frac{P_{102}P_{200} - 2P_{101}P_{201}}{P_{200}^2}$$

This may be used when the motion of a trough line is to be calculated over a considerable period of time, and under conditions where the trough is being accelerated or retarded. After the velocity and acceleration of the trough have been calculated, its displacement after time, t, may be obtained from the following formula:

$$(3) \qquad S_L = V_L t + \tfrac{1}{2}A_L t^2$$

The displacement as calculated from this formula, although it considers only instantaneous values of the velocity and acceleration, is very satisfactory in most cases for time intervals up to twenty-four hours. If the acceleration term is neglected, however, the inaccuracies after twelve hours may be considerable.

The evaluation of the movement of a trough line is demonstrated in figure 118. This calculation may be performed from information obtained directly from the weather chart after the isobars and isallobars have been constructed, using the formulas given on page 373. The first step is to draw on a separate piece of paper a line corresponding to the axis of the trough and to construct a perpendicular to this at the point where the movement is to be de-

termined. Along this perpendicular axis the pressure profile and the tendency profile are then constructed. If it is desired to obtain the acceleration of the trough line it will also be necessary to draw the tendency profile for the preceding weather chart in order to obtain the value for P_{102}. On the two perpendicular axes the points (1,0), (½,0), (0,0), (−½,0), (−1,0) are then indicated. The length of the units should be determined using the suggestions given on pages 372-373.

In the example given above:

$$V_L = -\frac{P_{101}}{P_{200}} \tag{4}$$

$$A_L = -\frac{P_{102}P_{200} - 2P_{101}P_{201}}{P_{200}^2} \tag{5}$$

$$S = V_L t + \tfrac{1}{2}A_L t^2 \tag{6}$$

Velocity of Pressure Centers—The velocity for the pressure center is determined in the same manner as that for the trough line but it is necessary to calculate the movement of the center along two axes in order to obtain its future position. The following formulas may be used for determining velocities:

$$V_{cx} = -\frac{P_{101}}{P_{200}}$$

$$V_{cy} = -\frac{P_{011}}{P_{020}}$$

Acceleration of Pressure Centers—The acceleration of pressure centers may be determined by using the same general formula that was used for trough lines. It will be necessary here, as with the velocity of the pressure center, to determine its acceleration along two axes in order to obtain its future position.

$$A_{cx} = -\frac{P_{102}P_{200} - 2P_{101}P_{201}}{P_{200}^2}$$

$$A_{cy} = -\frac{P_{012}P_{020} - P_{011}P_{021}}{P_{020}^2}$$

Movement of Pressure Centers—The movement of pressure centers may be calculated by the use of the ordinary formula for the motion of a particle, thus:

$$S_c = V_c t + \tfrac{1}{2}A_c t^2$$

Path of Pressure Centers—The path of pressure centers may be determined by calculating the movement along the x- and y-axes. Round centers move in the direction of the isallobaric ascendant. Oblong centers move along the longest axis, or between this axis and the direction of the isallobaric ascendant.

The velocity of a *round center* is given by:

$$V_c = \frac{I}{P_{200}}$$

Where I is the isallobaric ascendant. Thus, the velocity of a round center is directly proportional to the isallobaric ascendant and inversely proportional to the curvature of the pressure profile (P_{200}).

The path of a round center is given approximately:

$$\tan \theta = \frac{P_{011}}{P_{101}}$$

where θ is the angle between the x-axis and the path of the center.

MOVEMENTS OF FRONTS

The velocity and acceleration of fronts may be determined in exactly the same manner as for troughs. Strictly, the frontal surface represents a discontinuity and the following equation should be used.

$$V_f = -\frac{\dfrac{\partial p_1}{\partial t} - \dfrac{\partial p_2}{\partial t}}{\dfrac{\partial p_1}{\partial x} - \dfrac{\partial p_2}{\partial x}}$$

where p_1, and p_2 are the pressures at two points on either side of the front. For practical purposes, however, the ordinary trough formulas yield equivalent results, since finite differences replace the differentials in either case when the formulas for numerical differentiation are used.

The equation above yields some interesting facts governing the movements of fronts, although it is not employed practically. It may be seen at once that the velocity is *directly* proportional to the tendency difference across the front

$$\left(\frac{\partial p_1}{\partial t} - \frac{\partial p_2}{\partial t}\right).$$

Thus, the greater the tendency difference across the front, the more rapidly it will move. Also and this is very important, *if there is no tendency difference, the front will remain stationary.* The velocity is also *inversely* proportional to the difference in pressure *gradient* across the front

$$\left(\frac{\partial p_1}{\partial x} - \frac{\partial p_2}{\partial x}\right).$$

With poorly defined troughs, the front will therefore move rapidly. With sharp, or V-shaped troughs, the front will move slowly.

TYPES OF WARM SECTOR CYCLONES

Petterssen has distinguished several different types of warm sector cyclones, each of which shows peculiarities that are useful in determining its development. In Type "A," which is shown in figure 119-a it will be noted that the pressure is symmetrically distributed and that the isobars of the cold sector are curved anticyclonically. This type of cyclone is characterized by:

(1) Very slow occlusion,
(2) Usually moving as a stable wave without change in structure,
(3) Retardation of the lower portion of the cold front.

Type "B," as shown in figure 119-b, is characterized by a symmetrical distribution of the pressure and a *cyclonic* curvature of the cold isobars near the fronts. This type is characterized by:

(1) A rapid occlusion,
(2) Acceleration of the cold front, together with retardation of the warm front.

Type "C." This type is shown in figure 119-c, in which it is seen that the cold sector isobars near the cold front are curved cyclonically while those near the warm front are curved anticyclonically. It is characterized by:

(1) A moderate rate of occlusion, intermediate between types, "A" and "B,"
(2) Tendency toward rapid deepening.

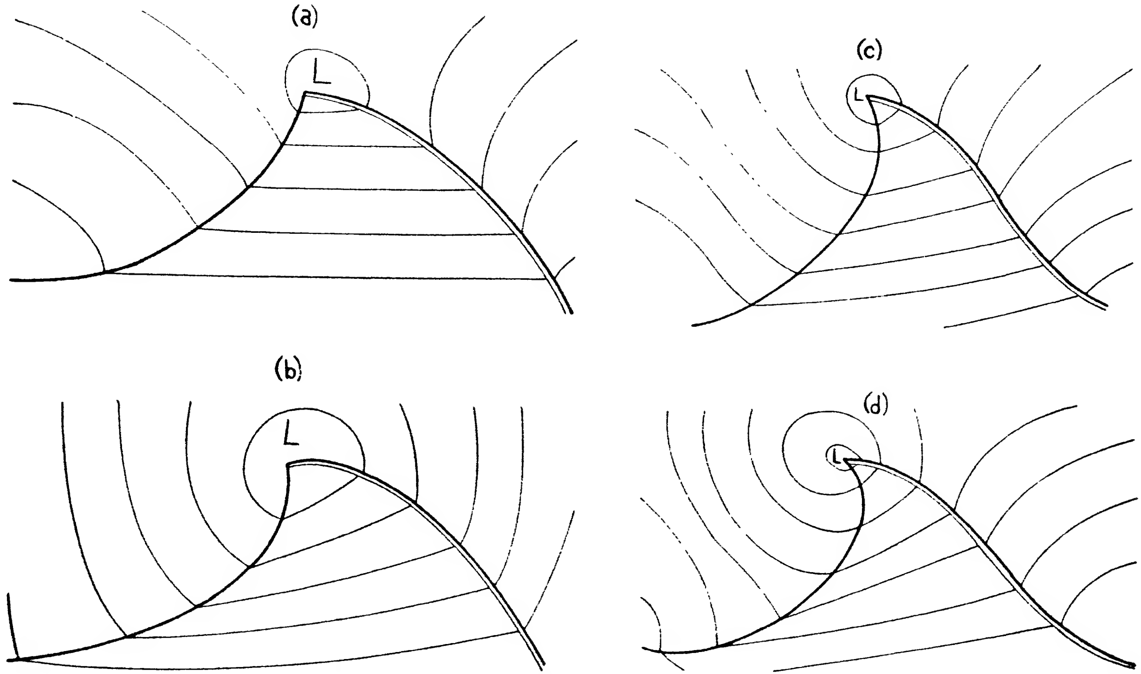

FIGURE 119.—VARIOUS TYPES OF WARM SECTOR CYCLONES

(*a*) Type "A." Characterized by symmetrical distribution of isobars, and their anticyclonic curvature in the cold sector. (*b*) Type "B." Characterized by symmetrical pressure distribution and cyclonic curvature of cold sector isobars. (*c*) Type "C" Characterized by cyclonic curvature of the cold sector isobars near the *cold* front and anticyclonic curvature near the *warm* front. (*d*) Type "D." Most common case. Northern portions of both cold and warm fronts are of type "B," and southern portion, type "A."

Type "D." Most commonly encountered are cyclones which are combinations of types "A" and "B." Especially frequent is the case where the northern portions of both the warm and cold fronts are of type "B," while the southern portions are type "A" (figure

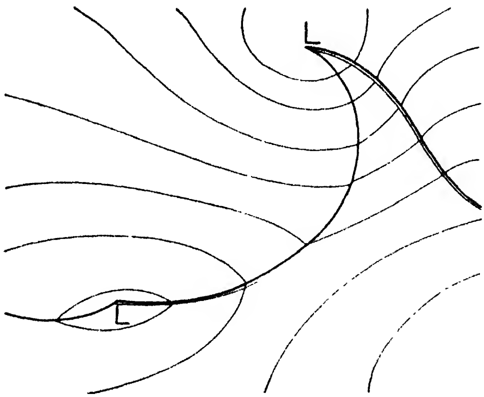

FIGURE 120.—FORMATION OF SECONDARY WAVE FROM TYPE "D" WARM SECTOR CYCLONE

The wave develops on the retarded portion of the cold front.

119-d). Generally, "B" tends to predominate on the cold front giving rise to Petterssen's type "D." This is perhaps the most frequent type of cyclone. It is characterized by rapid occlusion near the center, with a portion of the warm sector however, remaining open near the southern part of the cyclone. At the same time the lower portion of the cold front is retarded, and becomes subject to frontogenesis. This condition is very common in the case of *cyclone families,* and it may lead to the production of new waves to the rear of those already formed (figure 120).

DEEPENING AND FILLING OF PRESSURE CENTERS

The expressions "deepening" and "filling," when applied to moving pressure centers, refer to the pressure and the center of the *moving system,* and thus equations must be used which refer to the moving system, and not to individual stations on the earth. The

general formula for the deepening or filling of a pressure center is given by the following formula.

$$\Delta P = T^{0,0}t + \tfrac{1}{2}\left(\Delta T^{0,0} - \frac{P_{101}{}^2}{P_{200}} - \frac{P_{011}{}^2}{P_{020}}\right)t^2$$

where ΔP is the change in central pressure, $T^{0,0}$ the tendency at the center, and the other quantities have the same meaning as before. This may be seen to be essentially similar to the usual equation for motion, $S = V_0t + \frac{1}{2}\,a_0t^2$. If the *rate* of deepening or filling is constant, the "acceleration" term

$$\tfrac{1}{2}\left(\Delta T^{0,0} - \frac{P_{101}{}^2}{P_{200}} - \frac{P_{011}{}^2}{P_{020}}\right)t^2,$$

may then be dropped. This is permissible in nearly all cases for short time intervals, and for moderately long intervals where the development is proceeding in a uniform manner. *In such cases the tendency at the center gives the deepening or filling with sufficient accuracy for most practical purposes.*

The quantities included in this formula are those used in the equations for computing the velocity and acceleration of pressure centers, and thus the deepening may be computed with little additional labor. It is generally sufficient to compute the deepening or filling at the center and at one point on each of the coordinate axes. After this is done, and the future position of the pressure center is determined, new isobars for the future pressure system may be drawn.

Practical Applications of the Above Calculations—It might appear that the work necessary to carry on calculations similar to those outlined above for the paths of pressure centers and fronts, and for the deepening and filling of pressure systems would be very great. Actually, however, the necessary computation in most cases can be carried on in less than an hour. Generally, it will suffice to determine the movement of a single pressure center, and perhaps to estimate the deepening or filling of only a portion of a system. The increase in accuracy of forecasts made possible by these calculations is very considerable, and is generally well worth the time spent.

Each individual synoptic situation must be treated separately. In some cases it will be found desirable to calculate only the movement of fronts, while in other cases it may be necessary to compute

the path and determine the deepening of a pressure center. The accuracy which may be expected varies greatly with the local situation. In mountainous regions, where both the pressure and tendency fields are subject to considerable error, and where the movement of fronts is frequently greatly affected by the topography, the results obtained are frequently rather disappointing. In level country, however, and over the ocean, the results obtained in this manner are often remarkably successful.

In any event, the careful analysis of a number of situations by the use of the above formulas invariably will lead to a better understanding of synoptic developments. This knowledge is of great value in making merely purely qualitative estimates of weather developments. It is particularly desirable to construct the tendency profile of cyclones whenever possible, since the information obtained in this manner may be utilized in a number of ways in making qualitative estimates, as indicated by the forecasting rules on pages 382-385.

SUMMARY OF PETTERSSEN'S NOTATION

P_{000} Isobar

P_{100} Change in pressure per unit length along the x-axis. (Pressure ascendant along x-axis)

P_{010} Change in pressure per unit length along the y-axis. (Pressure ascendant along y-axis)

P_{001} Change in pressure per unit time. (Tendency)

P_{101} Change in pressure per unit length along the x-axis per unit time. (Isallobaric ascendant along the x-axis)

P_{011} Change in pressure per unit length along the y-axis per unit time. (Isallobaric ascendant along the y-axis)

P_{200} Curvature of pressure profile along x-axis.

P_{020} Curvature of pressure profile along y-axis.

P_{102} Change in tendency variation per unit length along x-axis.

P_{012} Change in tendency variation per unit length along y-axis.

P_{201} Change in isallobaric ascendant per unit length along x-axis.

P_{021} Change in isallobaric ascendant per unit length along y-axis.

SUMMARY OF PRESSURE AND TENDENCY FORMULAS

VELOCITY

Isobar $$V_i = -Th$$

where T is the local tendency, and h the distance between unit isobars.

Isallobar $$V_t = -\frac{\partial T}{\partial t}H,$$

where $\frac{\partial T}{\partial t}$ is the rate of change of the tendency, and H the distance between unit isallobars.

Trough or Wedge $$V_L = -\frac{P_{101}}{P_{200}}$$

LOW center or HIGH center $$V_{cx} = -\frac{P_{101}}{P_{200}}, \quad V_{cy} = -\frac{P_{011}}{P_{020}}$$

Front $$V_f = -\frac{P_{101}}{P_{200}}$$

ACCELERATION

Isobar $$A_i = \frac{P_{002} + 2V_iP_{101} + V_i^2P_{200}}{P_{100}}$$

Trough or Wedge $$A_L = -\frac{P_{102}P_{200} - 2P_{101}P_{201}}{P_{200}^2}$$

LOW center or HIGH center

$$A_{cx} = -\frac{P_{102}P_{200} - 2P_{101}P_{201}}{P_{200}^2}; \quad A_{cy} = -\frac{P_{012}P_{020} - 2P_{011}P_{021}}{P_{020}^2}$$

Front $$A_f = -\frac{P_{102}P_{200} - 2P_{101}P_{201}}{P_{200}^2}$$

GENERAL FORECASTING RULES

A number of very useful rules for forecasting may be derived from the quantitative formulas given above, even though the actual numerical calculations are not carried out. These rules all have a direct basis in the kinematical and dynamical formulas brought out above. They are thus not to be confused with purely empirical rules which have been derived from forecasting experience, and which have

no basis in kinematical theory. Even though time may not be available for performing the numerical calculations necessary to determine the movements of the various elements of a synoptic situation, nevertheless the application of these rules will aid greatly in preparing forecasts. These rules will be divided into several groups, each of them relating to a certain element of the synoptic situation.

PRESSURE CENTER MOVEMENTS

1. Circular centers of LOWS move in the direction of the *isallobaric gradient.*
2. Circular centers of HIGHS move in the direction of the *isallobaric ascendant.*
3. Moderately elongated centers move in a direction intermediate between the *isallobaric gradient* or *ascendant* and the longer axis. The more elongated the center the nearer to the long axis will be its direction of movement.
4. Very elongated centers move along the longer axis.
5. Either a cyclonic or an anticyclonic center moves *normal* to the isallobar running through the center.
6. The velocity of a pressure center is directly proportional to the *isallobaric gradient.* It is important to notice here that the velocity is proportional to the isallobaric *gradient* and not to the magnitude of the tendency. Frequently, the velocity is over-estimated in cases where large tendencies appear, but where the tendency *gradient* is not excessive. Pressure centers will remain stationary when the tendency is uniform in all directions.
7. The velocity is inversely proportional to the steepness of the pressure profile. Thus, pressure centers with steep profiles move slowly and those with flat profiles, if they are accompanied by isallobaric gradients of considerable magnitude, move rapidly.
8. A cyclonic center will be accelerated when the tendency profile is curved anticyclonically and retarded when it is curved cyclonically.

9. An anticyclonic center will be accelerated when the tendency profile is curved cyclonically and retarded when it is curved anticyclonically.
10. Circular pressure centers may be accelerated in any direction. Thus they may take considerably curved paths.
11. Very elongated centers are accelerated chiefly in the direction of the long axis and generally move in relatively straight paths.

DEEPENING OR FILLING OF PRESSURE SYSTEMS

1. The rate of deepening or filling of a pressure center is equal to the barometric tendency in the center.
2. A cyclonic center *deepens* when the line of zero pressure change (zero isallobar) lies to the *rear* of the line passing through the center normal to the direction of the movement. Conversely, the center *fills* when the zero isallobar lies *in front of this line.* (Figure 116.)
3. An anticyclonic center *increases* when the zero isallobar lies to the *rear* of the line passing through the center normal to its direction of movement. Conversely, the center *decreases* when the zero isallobar lies *in front* of it.
4. The deepening of the warm sector of a cyclone is equal to the tendency of the warm sector.
5. The deepening of the cold air near the cold front is approximately equal to the deepening within the warm sector.
6. Warm sector cyclones deepen with a uniform velocity.
7. The rate of deepening of the cold air along the cold front is constant.
8. Warm sector cyclones deepen with a constant velocity which is equal to the tendency at the top of the warm sector. This holds true until the occlusion process has become complete. After this, the cyclone fills at a somewhat slower rate.
9. A cyclonic center *increases* in strength when the tendency profile is curved *cyclonically*. It *decreases* when it is curved *anticyclonically*. The magnitude of the curva-

ture of the tendency profile determines the amount of increase or decrease in strength.

10. An anticyclonic center *increases* in strength when the tendency is curved *anticyclonically*. It *decreases* when it is curved *cyclonically*. The curvature of the tendency profile determines the increase or decrease in strength.

MOVEMENT OF FRONTS AND FRONTOGENESIS

1. The velocity of a front is *directly* proportional to the tendency differences across the front. It is *inversely* proportional to the pressure gradient difference across the front. Of these two criteria for the movement of a front, the first is the more important and it provides a very useful means for a quick estimation of the frontal velocity. It should be noted that even with no pressure gradient, a front may still move if a tendency difference exists across it. It is not true, therefore, that a front will become stationary if the isobars on the two sides are parallel to it. It will become stationary if the tendency difference also disappears.
2. The occlusion of symmetrical cyclones is directly proportional to the warm sector tendency. No occlusion will take place when the warm sector tendency is zero.
3. Occlusion of a cyclone is directly proportional to the deepening of the pressure system.
4. *Frontogenesis* will occur when the front is subjected to ***retardation.***
5. *Frontolysis* will occur when the front is subjected to ***acceleration.***
6. The portion of a cold front at a considerable distance from the center is generally retarded and thus exposed to frontogenesis.
7. Stationary fronts are subject to frontogenesis when the isallobaric gradients are directed toward the front. Similarly they are exposed to frontolysis when these gradients are directed away from the front.
8. Waves are most readily formed along the boundary between two parallel currents moving in the same direction.

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Petterssen summarizes his various rules concerning the development of pressure fields. He gives many useful rules for practical use in all types of synoptic situations.

3. Weather Analysis and Forecasting: Sverre Petterssen. New York. 1940.

An excellent advanced treatise on the various techniques employed in modern weather analysis.

CHAPTER 20

FORECASTING FROM LOCAL INDICATIONS

INTRODUCTION

It is desirable, wherever possible, to employ sufficient data to construct a weather map covering a considerable area surrounding the region for which forecasts are to be made. There are many cases however, where this is not practicable. In such situations forecasts must be prepared from purely local indications. In any event, it is very desirable for the forecaster to recognize the significance of such meteorological information as may be available locally. Any forecasts based on such data must, necessarily, be for relatively short periods of time and subject to considerable inaccuracy.

INDICATIONS OF CLOUDS

The type and direction of movement of clouds is one of the most important elements to be considered in making any local forecast. Cirrus clouds, the highest of all, often give indications of approaching stormy weather many hours in advance of its arrival. In this case it is necessary to observe the changes that take place in the cirri as well as their general form, for this has a considerable bearing on the type of weather that may be expected. The time of year during which these observations are made also is a very important factor. Cirrus clouds which appear in advance of an approaching warm front invariably show a marked arrangement in parallel bands (see figure 94) and a gradual increase in amount so that within the space of a few hours a marked increase in the amount of sky covered may be noted. If the clouds are not arranged in bands and if they show no marked increase, very little significance may ordinarily be attached to them.

As the cirrus clouds thicken and increase in amount it will frequently be found that an altostratus system will appear. This type of cloud is a very definite sign of an approaching warm front

or occlusion, as the case may be, and generally indicates that stormy weather may be expected within a very few hours.

Cumulus type clouds are of varied significance. Generally the appearance of cumulus clouds early in the morning during the summer indicates that showers may be expected during the afternoon. If, on the other hand, the first cumulus clouds do not appear until late in the morning the shower activity, if any, will be very light and will not occur until late in the day. The form of cumulus clouds is of considerable importance in determining the future weather. If the vertical extent of the clouds is rather limited and their tops are comparatively flat (see figure 101) very little activity may be expected. If, on the other hand, the cumulus clouds bulge upward freely and show no evidence of horizontal stratification in their upper levels (see figure 102) it is evidence that considerable cumulus activity will occur. The so-called "fair weather cumulus," are of the type first mentioned above, with comparatively little vertical extent and a notable tendency toward horizontal stratification in their upper portions.

With all types of cumulus activity, rainfall rarely commences before the upper portions of the clouds have reached the freezing level. This stage in the cumulus development is shown by the formation of a hazy mantle around the upper portions of the cloud as the *ice crystal stage* is reached (see figure 102). Thus, although many cumulus clouds may be present in a region, unless some of them exhibit signs of this ice crystal stage, no shower activity need be expected.

WIND

The direction and velocity of the wind in the temperate latitudes is often a fairly reliable indication of coming weather changes. Thus, the tendency of the wind to back into southerly directions and increase in velocity is very frequently a sign of an approaching cold front. This criterion is especially valuable in the case of cold fronts since no prefrontal cloud systems appear with them. Accompanying this shift of the wind toward the south (in northern latitudes) an increase in both temperature and humidity will occur. An indication of the intensity of the coming cold front can frequently be obtained from the velocity of the wind and the temperature and humidity of the air.

Other useful indications may be obtained from the wind direction in individual cases. Thus, it is generally found that the occurrence of a northeast wind along the New England coast precedes a condition of low clouds and misting rain very characteristic of weather in that region in certain seasons. Similarly, a persistent southerly wind over the Gulf coast region generally precedes the appearance of a low status fog characteristic of invasions of tropical air in this region.

Sudden shifts in the wind, which generally indicate the passage of fronts, are almost always followed within a comparatively short time by improvement in the weather. In the northern latitudes the wind generally shifts from the south or southwest to north or northwest with the passage of cold fronts, and from the southeast or south to the south or southwest with the passage of warm fronts. Since the weather within the air mass following such a wind shift depends greatly upon the characteristics of the air, it is not to be expected that any accurate forecast of coming weather conditions can be expected without a rather complete knowledge of the air mass properties.

TEMPERATURE

A gradual increase in temperature accompanying southerly winds which precede the appearance of a front is generally an indication of stormy weather to come. Many times, however, this increase in temperature may simply be due to gradual radiational heating within an air mass behind a front and, therefore, this fact alone without regard to the direction of the wind may be misleading. Sudden rapid falls in temperature following a cold front passage are generally recognized easily and in most cases can be interpreted as indicating clearing weather. Whether or not this clearing will take place slowly or rapidly depends on the air mass characteristics and also to a great extent on the location of the observation station. This latter point is of considerable importance in many cases. For example, the passage of a cold front over a station such as St. Louis may be followed by rapidly improving weather, whereas, the passage of the same cold front over Cleveland, Ohio, with its lake environment may be followed by continued snow flurries and bad weather for a considerable period of time. Furthermore, in many cases the passage of a cold front is followed by convergent

activity that may delay clearing for a long period. It is thus clear that forecasts based on purely local indications are subject to great errors in many cases.

ATMOSPHERIC PRESSURE

The use of a barometer to indicate atmospheric pressure in connection with the other elements such as temperature, wind direction and cloudiness mentioned above will increase to a considerable extent the probable accuracy of local forecasts. In general a decrease in atmospheric pressure indicates the approach of stormy weather and an increase in pressure denotes generally clearing weather. This very generalized rule is subject to many modifications, however, depending on the synoptic situation. Unless the pressure decreases at a fairly rapid rate it cannot be expected to indicate very reliably the approach of a storm area, since the decrease may be due to a widespread effect not connected with any storm in the immediate vicinity. Care must be taken in any event to allow for the average diurnal change in pressure and not to confuse this with changes due to approaching storms. If the pressure increases rapidly after the passage of a cold front, conditions may be expected to improve quite rapidly, whereas if the increase in pressure takes place slowly it may be an indication of the onset of convergent conditions with continued poor weather.

THUNDERSTORMS

Air mass and cold front thunderstorms generally give considerable advance local indication of their appearance. This is particularly true of air mass thunderstorms since the air mass properties which are necessary for their formation may frequently be recognized by purely local observations. Thus, highly humid warm weather, where the dew point exceeds 60° F. and the temperature exceeds 80° F. during the mid-forenoon, together with fairly light surface winds are almost invariably indications of thunderstorm conditions in the Gulf coast region of the United States. This type of weather is further associated with the early formation of cumulus clouds which show marked development as early as noon.

Generally in this connection it will be noted that the cumulus clouds do not exhibit any stratification but rather show the towering form characteristic of the early stages of cumulonimbus clouds, even as early as the middle or latter part of the forenoon.

In other parts of the country somewhat different indications are characteristic precursors of thunderstorms. It has been noted, for instance, in the New Jersey region that air mass thunderstorms are generally preceded by a rather thick haze at the surface. If this is accompanied by rather high humidity and temperature, as well as by notable cumulus development during the late forenoon, thunderstorm activity almost invariably occurs during the afternoon or early evening. The haze mentioned in this regard is probably the typical tropical air haze characteristic of invasions of this region by tropical maritime air masses and has been mentioned by observers from many parts of the world in this connection. It is a different type of haze from that due purely to smoke pollution and is distinguishable from the latter by the opalescence and bluish color which it imparts to distant objects as contrasted with the dull gray appearance of smoke haze which is produced in the neighborhood of large cities.

The type of cumulus cloud development which is observed in the latter part of the forenoon is a useful criterion for thunderstorm activity in almost any locality. If marked cumulus development occurs early in the forenoon with the tops of the individual cumuli building up into towering summits which are not limited by inversion levels, thunderstorm activity may well be expected during the afternoon or evening. On the other hand if the cumulus clouds which appear during the morning or early afternoon show comparatively little vertical development and are limited by inversion levels the chances of thunderstorm activity are very slight. The appearance of the cloud type, *altocumulus castellatus,* at any time during the late morning or afternoon is a particularly valuable indication of thunderstorm activity since this cloud type indicates rather marked instability in the intermediate cloud levels. Care must be taken in identifying this type of cloud and in not confusing it with isolated cloudlets of the altocumulus type which may appear to be arranged along a line suggesting the *castellatus* cloud. The true *castellatus* consists of a long, even-based cloud at the altocumulus level with a number of pronounced bulges, giving it its crenellated appearance (figure 98). This type of cloud generally pre-

cedes the actual appearance of thunderstorms by several hours and, therefore, gives considerable advance warning of their appearance.

The cloud type, *mammatocumulus* (see figure 121), is an excellent indication also of thunderstorm activity but it generally is not formed until an active cumulonimbus cloud has been produced and, therefore, it gives little warning of the appearance of thunderstorms. The remarkable structure of this type of cloud is probably due to the downward currents of cold air which are found in the lower

FIGURE 121.—MAMMATOCUMULUS CLOUDS

This cloud type indicates marked instability and generally accompanies thunderstorm activity.

parts of cumulus clouds. When a number of these downdrafts are present in a single cloud they may cause the lower surface to bulge downward in the typical mammatus form. Severe turbulence and hail may be expected under these conditions, and this cloud type should be carefully avoided by pilots for this reason.

It is common experience that thunderstorm activity is rarely, if ever, observed in connection with strong upper winds. It can be stated almost as a rule that if the upper winds exceed a velocity of 30 to 40 miles per hour at the levels in which cumulus clouds

form, viz., 4000 to 15,000 feet, that marked cumulus activity will not be experienced. This is undoubtedly due to the fact that individual vertical currents which are vitally necessary for the formation of cumulus clouds can not be maintained under these circumstances. Furthermore, the temperature lapse-rate in the atmosphere tends to be stabilized due to the mixing which is produced by these high wind velocities.

The barometer almost invariably falls irregularly one or two hours prior to the occurrence of a thunderstorm. This is best shown by a micro-barograph, where these minor fluctuations, amounting to only a few hundredths of an inch, appear clearly. If this falling of the barometer occurs with the general thunderstorm indications provided by cumulus clouds, the chances of a storm passing over the station are rather high. The occurrence of the usual cumulus activity with no barometric fall indicates thunderstorms in the general region but not at the station. With very small storms the barometric fluctuation may be too small to observe.

FOG

The forecasting of fog, although connected in many ways with the general synoptic situation, depends to a great extent on entirely local phenomena. For this reason its successful forecasting demands a close study of various purely local effects, including topography, normal direction of the surface winds under conditions of very weak pressure gradient, sources of atmospheric pollution, effects of neighboring water bodies, etc. Each of these factors must be studied in considerable detail in connection with various types of fog and it is only through rather wide experience that the forecaster will have any degree of success with this extremely difficult problem.

One of the first things that should be investigated is the general subject of wind direction and velocity under conditions of poorly defined pressure gradient. It will generally be found that dense ground fog will not form in conditions of complete calm since turbulent mixing of the lower few hundred feet of air is necessary if the saturated stratum is to exceed a few feet in thickness. A condition of complete calm for several hours is generally possible only in flat regions, however, and in rolling country drainage of cold air into regions of low elevation usually produces light winds and suf-

ficient mixing to cause fog formation. The direction and velocity of these light "gravity," winds is very important to the forecaster and should be known for each locality. It is also generally found, at least in inland regions, that ground fog is not produced if the wind velocity is over 6–8 miles per hour. This rule does not hold true along the coast where the sea breeze frequently amounts to 15–25 miles per hour and may bring inland with it a low stratus deck or surface fog. Prefrontal fog along coastal regions is also generally associated with moderate wind velocities.

In inland regions fog is generally accompanied by low wind velocities so it occurs in the majority of cases in connection with an anticyclone thus, abnormal barometric pressure is some indication of fog possibility.

Since a fairly high concentration of nuclei of condensation is required to produce dense fog, it is generally observed that unless a well marked haze exists during the afternoon and early evening that fog will be unlikely later that night. It is generally found in this regard that if conditions of temperature and humidity are favorable for fog, with insufficient nuclei of condensation, however, that either the fog which is produced will be very light or no fog at all will occur and only a heavy dew will be formed.

The direction of the wind in nearly every locality is an important indication of fog probability. This is particularly true in regions which are comparatively close to important bodies of water or to large cities. It is commonly observed that the ground fog produced along large rivers, which is generally of the "steam mist" variety, will affect only a comparatively small territory immediately adjacent to the river, so that unless the wind blows from a certain direction, neighboring areas will not be affected. The importance of atmospheric pollution for fog formation is also shown by a study of the effect of wind direction on fog frequency in the neighborhood of large industrial centers. Many observers have noted that unless the wind is of such a direction as to bring smoke polluted air from the city over a region, dense fog is very unlikely.

The temperature distribution in the upper atmosphere may frequently be obtained without difficulty by airplane pilots. This information is invaluable in preparing fog forecasts since it gives very useful indications of atmospheric stability. If temperature soundings obtained in this manner indicate marked stability with isothermal layers or levels of inversion, the conditions are much more

favorable for fog formation than if a fairly steep lapse rate is present. Furthermore, information concerning the inversion levels obtained in this manner give useful indications of the elevation at which stratus clouds will generally form. Temperature data obtained in this manner are of use primarily along coastal regions or in areas directly affected by fresh maritime air, for it is in this air mass type that stratus fog occurs. Since high fogs are practically unknown in the interior of the United States this information is of little value there except as it indicates the presence of a surface inversion with the obvious implication of ground fog formation.

The appearance of beams of light at night is a rather useful indication of conditions favorable to fog formation. If the path of a powerful beam of light such as that projected from an automobile headlight is well defined, the presence of abundant nuclei for fog formation is well established. This phenomenon is even better shown by vertically directed light beams such as those thrown by ceiling light projectors used for making aeronautical observations. When abundant nuclei are present the path of the ceiling light can be distinctly observed, especially when it is viewed from the immediate vicinity of the projector.

The effects of dynamic heating on air which descends from higher levels should be given careful consideration in evaluating the influence of winds from various directions on local fog situations. These foehn heated winds, although they may appear of very little significance in other respects, may be of considerable importance in their effect on fog formation. Thus it frequently will be found that winds from a certain direction, which have undergone dynamic heating, may be very effective in preventing fog from forming or in dissipating fog already formed. This factor should be given careful consideration in all regions of irregular terrain. It is of particular importance in inland valleys situated several miles from the seacoast. In such situations fog formation may be very closely connected with wind direction, for one direction may bring fog-producing maritime air into the region while another direction may cause its dissipation by bringing warm, foehn heated air from the interior.

It has been found by practically all of those who have studied the general subject of low fog that it will not be formed unless the temperature and the dew point are comparatively close together during the early evening. The well known "fog graph" of G. I. Taylor is an instance of this means of studying the situation. Taylor simply

plotted on a graph having as coordinates the dew point depression and the air temperature, at a certain time in the early evening, all situations which seemed to be favorable for fog formation. He found that he was able to draw a line separating portions of the field in which fog might be expected, and in which it would not be expected. This type of study is generally rather inconclusive and applies only to very definite situations. Actually, it simply shows a relationship between dew point depression and radiational cooling.

If the rate of cooling of the surface air can be determined, as may frequently be done by a consideration of an earlier situation involving similar air, the lowest point to which the temperature will fall during the night may be fairly well established. If this is several degrees above the dew point, as measured in the early evening, it will be unlikely that fog will form. This depression of the dew point, as observed in the early evening, is useful only in very definite cases. In indefinite situations it is unreliable.

CHAPTER 21

AIRLINE METEOROLOGICAL DEPARTMENTS

INTRODUCTION

The meteorological organizations of scheduled airlines differ greatly according to the needs of the various companies. Small companies may find it unnecessary to maintain a weather organization of their own at all. These companies often find that the weather information furnished by governmental agencies is quite adequate for their needs. Larger companies, on the other hand, may feel the necessity of supplementing the information obtained from governmental sources by more detailed information from their own meteorological staffs. The present chapter will point out briefly some of the ways in which a company meteorological staff may be of use to the company. It will also indicate some of the types of company meteorological organizations that are in successful use.

Two quite different types of meteorological organizations are ordinarily employed. In one type, the company meteorologists are primarily forecasters, who do very little observational work. This is the situation in the United States, where the United States Weather Bureau is practically the sole agency for making observations. In the other type of organization, the company meteorologists may spend much of their time taking observations or training other company employees in observational technique. This is the situation with many international airlines, where practically the only sources of weather information may be the observation stations of the company.

METEOROLOGICAL ORGANIZATIONS OF UNITED STATES AIRLINES

Meteorologists with United States airlines serve primarily as forecasters. The observational work is performed almost exclusively by the U. S. Weather Bureau, and the data obtained in this way are transmitted by means of the communications system of the Civil Aeronautics Board.

The meteorologists in most instances perform various other duties in addition to actual forecasting. This is especially true at the less busy terminals, where they may perform a certain amount of routine clerical work. In the busier terminals, though, practically all of their time is consumed in the preparation of forecasts, and in discussing the weather situation with pilots and flight dispatchers.

This informal discussion of the weather situation with flying and supervisory personnel is perhaps the most valuable part of a meteorologist's services. Much of the success of the individual meteorologist, in fact, depends on his ability to instill confidence in his forecasting ability in the pilots and flight dispatchers with whom he deals. Since this is the case, the personality of the airline meteorologist is of no small importance. A man who finds difficulty in expressing himself, or who is a poor "mixer" with other people, will rarely be successful in airline meteorology, no matter how skilled he may be in technical knowledge.

In some companies, flight control responsibility is given to meteorologists who are well qualified and who have had a considerable amount of experience. This is a very satisfactory arrangement in cases where there is insufficient work to employ a man for each position. Since flight dispatching is very largely a meteorological problem, a successful meteorologist in many cases will be well qualified to assume the additional duties of the flight dispatcher.

In the larger companies the chief meteorologist is generally an important member of the operations staff. He has charge of the organization of the meteorological department itself and usually supervises meteorological instruction for pilots and flight dispatchers. He also deals with the various governmental agencies which are concerned with weather observations and their distribution.

The instruction of flying and supervisory personnel in meteorology is important if the fullest use is to be made of weather information made available by government and company meteorologists. Pilots and flight dispatchers should know how to interpret a weather map, and should have at least some knowledge of its construction. Pilots especially, should know a good deal about the physical processes of weather, so that they will be able to formulate their own forecasts from observations which they make while flying.

All operations personnel should understand clearly the limitations of modern forecasting methods. They should realize that some situations lend themselves to precise forecasts while others cannot

be interpreted in any but general terms. If this point is well understood, much misunderstanding can be avoided.

Instruction of operations personnel in meteorology may best be carried on by a combination of correspondence and classroom instruction. A series of about 20 lessons, mailed at weekly or bi-weekly intervals will generally suffice to present the fundamentals of meteorology. Since it is hardly possible to make such lessons sufficiently detailed to cover all points, they may be supplemented by reading in meteorological texts. Each of the lessons should include a few well chosen questions which the students will be expected to answer and return for correction and comment. Several classroom lectures should be delivered during the period covered by the correspondence course in order that difficult points may be clarified. During these classes many interesting points are brought out from the personal experiences of the pilots.

After the completion of the formal course, occasional meetings of pilots and flight dispatchers with meteorological personnel to discuss weather conditions should be continued. Pilots are in a unique position, in that members of no other occupation have as great an opportunity to study weather conditions at first hand. For this reason, meteorologists can invariably learn much of value from flying personnel in such informal discussions.

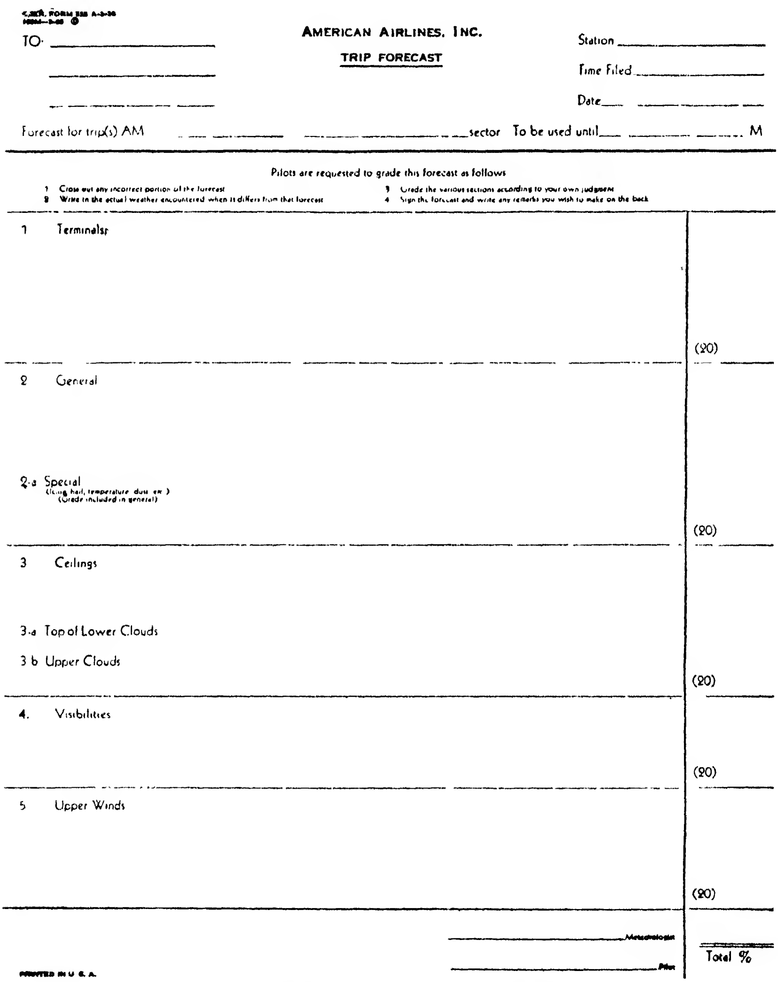
Members of the meteorological staff must be kept abreast of all developments in the field. The chief meteorologist should make it his duty to investigate all improvements in forecasting and observational technique, and to make them available for use by the company staff. Research by individual staff members is particularly desirable, since it keeps the spirit of investigation alive. In a science which moves as rapidly as meteorology, the lack of a spirit of research can only lead to rapid stagnation.

Types of Company Forecasts—Three general types of forecasts are commonly issued by airline meteorologists. The first of these is the *trip* or *route* forecast. This is prepared in considerable detail and is used by pilots and dispatchers in planning the disposition of individual flights. The *trip* forecast is prepared for each trip, while the *route* forecast is issued for a definite period of time for a certain route. The *route* forecast is often used where a large number of flights traverse a given route within a relatively short time. The *trip* forecast is used where but few flights operate over a given airway, or where the forecast must be especially detailed.

Since the route forecast generally refers to a forecast period of 6–12 hours, it can hardly be prepared in as much detail as a *trip* forecast which refers only to the time required by a single flight to travel 300–800 miles.

The trip or route forecast generally describes the weather conditions in several sections. A representative type of trip forecast is that prepared by American Airlines (see figure 122).

A rather general forecast, covering a period of 24–36 hours, is



AMERICAN AIRLINES, INC.

TRIP FORECAST

TO: ____________

Station ____________

Time Filed ____________

Date ____________

Forecast for trip(s) AM ____________ sector To be used until ____________ M

Pilots are requested to grade this forecast as follows

1 Cross out any incorrect portion of the forecast
2 Write in the actual weather encountered when it differs from that forecast
3 Grade the various sections according to your own judgment
4 Sign the forecast and write any remarks you wish to make on the back

1 Terminals (20)

2 General

2-a Special
(Icing, hail, temperature, dust, etc.)
(Grade included in general) (20)

3 Ceilings

3-a Top of Lower Clouds

3 b Upper Clouds (20)

4. Visibilities (20)

5 Upper Winds (20)

____________ Meteorologist

____________ Pilot

Total %

PRINTED IN U. S. A.

FIGURE 122.—SAMPLE AIRLINE TRIP FORECAST FORM

generally prepared for use by dispatching personnel in planning the operations for the following day. This *operations forecast* is of comparatively little practical importance, and is merely intended to present a general picture of weather conditions in the forecast area.

The *traffic forecast* is a comparatively recent development in airline meteorology. It is prepared for use by the reservations or traffic department in advising passengers of flying weather conditions. In the earlier development of aviation, little, if any, information regarding flying weather was made available for use by passengers. The recent, rapid development of passenger travel, however, has given rise to a persistent demand from the traveling public for reliable and up-to-date information regarding the flying weather.

This type of forecast generally specifies the probability of completion of each flight made by the company within the forecast area for a period of 12–24 hours. The forecast also includes a very brief summary of the weather which each flight will encounter. The *traffic forecast* must be in non-technical language, and worded in such a manner as not to alarm passengers who are unacquainted with the significance of various types of weather.

Since these forecasts are generally read directly to inquiring passengers by reservations clerks, considerable care must be used in the choice of terms. Thus, the word "squall," may cause considerable trepidation in a passenger who is making his first airplane trip. A "few scattered showers," on the other hand, would not worry him, yet might convey as accurate a picture of weather conditions. A sample traffic forecast is given below as it might be prepared for an airline operating between Chicago and New York.

"Probable short delays trips 7 and 8 between Buffalo and New York. Other trips normal. General light rain New York state. Remainder course scattered clouds to clear."

GENERAL METEOROLOGICAL ORGANIZATIONS

In regions where governmental agencies do not provide adequate observational material, the company must arrange for its own observations. The meteorological organization will approximate a complete weather bureau in such cases with observers, supervisors and forecasters. The observers may be company employees who are generally engaged in other work, such as radio operators. The

supervisors may be trained meteorologists who will spend most of their time in instructing observers, and maintaining or improving the observational technique. The forecasters should be well trained meteorologists who correlate the data obtained from the company observation stations and others, construct weather charts, and issue forecasts to flying and dispatching personnel.

The forecasts issued by the meteorological department in such cases are essentially the same as those issued by the various U. S. airlines. The difficulty of making detailed forecasts is considerably greater due to the usual lack of sufficient data.

The location of observational stations is very important, especially in situations where the total number of stations is small and each one must represent a large area. As a general rule, observation stations in rough topography should not be located on high peaks. Weather conditions reported from such locations will never represent the average of the region satisfactorily. Much more representative observations will be obtained from stations in broad valleys, where scud clouds which often cap the higher mountains, do not interfere with observations.

Each weather observation station should regularly observe all of the weather elements which affect flying, and also those which are necessary in the constructing of weather charts. These include: ceiling, visibility, sky condition, state of weather, temperature, dew point, wind direction and velocity, barometric pressure and pressure change. Cloud data are also useful, especially when the observation stations are widely separated. Sufficient pilot balloon observations should be available to present a complete picture of upper winds. This is especially necessary for over water flights or for instrument flights over land.

CHAPTER 22

FLIGHT DISPATCHING

INTRODUCTION

Flight dispatching is so intimately connected with aeronautical meteorology that a chapter on it seems to be required in a book on this general subject, although it is primarily a phase of airline operation. In the rapid growth of aviation, the need for flight control from the ground was recognized soon after the development of radio communication made it possible. As the range and speed of airplanes has increased, the duties and responsibilities of the flight control personnel have also increased. With flights of 600–800 miles a commonplace, and much longer flights in immediate prospect, the great importance of the flight dispatcher is easily recognized.

WEATHER LIMITATIONS

Certain restrictions as to what constitutes "flyable weather" are generally in effect for all scheduled airlines. These restrictions are decided upon in most cases by agreement between the governmental agencies which control aviation, and the individual companies. In cases where no aviation branch of the government exists, company regulations fix the flying limitations.

Weather limitations vary widely. They are influenced by: (a) the terrain, (b) whether flying is to be carried on both in daylight and darkness, or only in daylight, (c) the type of airplanes being operated, (d) the airway facilities, (e) the experience of the pilots, (f) the navigational aids available, including radio facilities, (g) the number of weather reporting stations. Each of these factors is of great importance and must be given careful consideration in establishing the weather limits below which flying cannot be permitted.

Limitations are of two general types: Those governing *en route flying*, and those governing *terminal conditions*. *En route* limitations

involve ceilings, visibilities, icing conditions and severe turbulence. *Terminal* limitations are concerned chiefly with ceilings and visibilities. When contact flying only is being carried on, the en route and terminal limitations will be identical. When instrument flying is possible, however, the en route limitations may concern only icing and turbulence. In this case, the terminal ceiling and visibility limitations are determined mainly by the terrain, and the radio aids available.

In rough terrain, particularly in regions where radio reception may be unreliable, the terminal limitations may require an unlimited ceiling with no more than scattered clouds in the sky. In level country, on the other hand, where radio reception is reliable and no obstructions are present, the terminal limitations may only require a 200–400 foot ceiling. With improvement in blind landing systems, it may be possible within a short time, to effect safe landings with even lower ceilings. In between the two extremes of terminal limitations mentioned above, occur every possible intermediate limitation.

En route weather limitations are similarly greatly influenced by the terrain. This is especially true for contact flying and to a considerably smaller extent, for instrument flying. Instrument flying is generally prohibited for single motor airplanes. Thus, en route limitations for single motor airplanes generally require contact flying. At the present time, very few single motor airplanes are used for scheduled passenger flights.

DISPATCHING JUDGMENT

Weather limitations represent the lowest possible allowable limits for flying. Since this is true, it is obvious that at times the poorest weather conditions in which flights may safely be operated, may be well above the official limitations. The flight dispatcher must exercise the most expert judgment at all times in weighing the various elements that affect the safety of flight. He must consider the present weather conditions as known from actual observations and deduced for intermediate points. He must decide what changes in the present weather conditions are likely to occur during the time the airplane will be in flight. He must consider carefully, pilot experience and ability, radio facilities, terrain, condition of the airways

(including landing fields, revolving beacons, and radio aids), and mechanical condition of the airplanes. A decision to allow a flight to depart or proceed must take into consideration all of these factors. If any one of them indicates that safety is being compromised, permission to proceed must never be granted. This is particularly true with regard to possible weather changes which may occur during the progress of the flight. Many weather situations do not admit of precise forecasts of weather elements affecting flying, even for relatively short intervals. In such cases, flight clearance must be made on the basis *that the most unfavorable conditions possible under the given weather situation will occur.*

This means that the individual weather reports from an area must not be accepted as sole guide to flight clearance. When a moderately active cold front moves across a region, for example, it may cause extremely varied weather conditions. All weather reporting stations in the area may report entirely favorable conditions, yet the territory between these reporting stations may be experiencing unfavorable weather conditions. Many other examples similar to this could be cited, but it will suffice to say that a careful study of the synoptic situation is as important as an examination of individual weather reports.

On the other hand, it should be realized that the construction of a weather chart, and its interpretation, are only means to an end. They are only useful as they assist in the preparation of accurate forecasts. They do not lead to quantitative forecasts of the accuracy demanded for airline operation, unless they are supplemented by frequent weather reports from the airway and its vicinity.

The successful dispatcher will thus closely correlate the regular airway reports with the synoptic weather chart. Neither of the sources of information is sufficient alone to provide all of the necessary weather information. Changes in airway weather reports often appear to be without any definite cause or relationship when viewed separately. They all appear as expressions of the general weather situation, however, when studied with the aid of a modern synoptic weather chart. Yet the weather chart can only indicate in a general way the direction and magnitude of weather trends. It must be supplemented by ample airway reports to give forecasts a strictly quantitative meaning, and to provide a continual check on them.

The dispatcher who watches only the airway reports, without

reference to a weather map, is in a perpetual quandary. He never knows what to expect next in the weather developments. Each change in the weather is a surprise. This type of dispatcher will enjoy but slight success compared to the one who understands the interpretation of weather maps and the physical processes of weather, and who makes a continual intensive study of the reason for all weather developments in his region. Even with staff meteorologists to construct weather maps and interpret the synoptic situation, flight dispatchers should understand all phases of aeronautical meteorology if they are to derive the greatest benefit from the meteorologists.

CHAPTER 23

CLIMATOLOGY

INTRODUCTION

The general subject of climatology as it affects aviation can hardly be discussed here except in the most general terms. Many climatological studies have been undertaken for various parts of the world and much of the information necessary for determining the suitability of various routes to air travel can be obtained from these publications. The most complete of such climatological studies for the entire world is the "Handbuch der Klimatologie" by Köppen and G. and R. Geiger. A number of more detailed studies of individual parts of the earth are also available and may be consulted in connection with individual problems.

LOCATION OF AIRWAYS

The factors which enter into the choice of air routes are those which affect flying in general, namely: (a) ceiling, (b) visibility, (c) squalliness, (d) icing conditions, (e) prevailing winds. Other climatic factors are of relatively minor importance insofar as the successful completion of flights is concerned. Some of the above data may be obtained from ordinary climatic surveys. Others, such as the ceiling and the visibility, can only be obtained through a special survey conducted from the aeronautical point of view.

In all such surveys the position of storm tracks during various seasons of the year should be known in order that the routing may be arranged to avoid severe storm activity during certain seasons of the year. It is often found desirable to change the routing of long airways so that the most direct route is employed during the season of favorable weather, and a somewhat longer route with better weather conditions, during the season of unfavorable weather.

In mountainous regions it will be found many times that relatively minor changes in an airway will result in very marked im-

provement in general weather conditions. Short airways which cross mountainous terrain, even those which are only a few hundred miles in length, will often benefit materially by a slight shifting of their course. Rather careful study of weather conditions prior to the establishing of airways in mountainous districts, is therefore of prime importance. As a general rule, the airway should be located to the lee of high elevations, giving due regard to the prevailing wind direction and average storm tracks. In such localities the air will have a downward component much of the time, with a resulting tendency toward clearing weather. The windward slopes of mountains should be avoided, since ascending currents with resulting bad weather will be found here. Only a careful survey extending through one or two winters will serve to locate the best airway in mountainous regions. A hastily installed airway, with no attention paid to choosing the most favorable location, is a very poor investment.

Along coastal regions the problem of fog is all-important. Here, it will almost invariably be found that it is desirable to locate the airway some distance inland. Especially is this true along coastal regions which are bordered by hills. In such cases, the incidence of fog a short distance inland may be much less than along the immediate coast. Here again the study of climatological records will be found to be very useful in planning the detailed location of airways.

In transoceanic flights the choice of a route almost invariably involves a compromise between a relatively short distance with poor weather, and a longer distance with better weather. The question of intermediate fueling stations is of great importance, and the weather of such intermediate stops is as important as that of the terminals. In the higher latitudes, the problem of fog in oceanic regions becomes of importance and may make it impossible to use certain terminals during the foggy portion of the year. This is especially true in the regions of the north Pacific and the western portion of the north Atlantic where the fog incidence is very high.

The velocity of upper winds may be of considerable importance in determining the feasibility of long over-water flights. This matter should be given careful study if these are contemplated. It may thus be found to be economically impossible to operate certain types of equipment on long flights if the average winds at flying levels exceed a certain value. This value should be determined from a

consideration of the fuel and pay-load characteristics of the available equipment.

The question of low ceilings and visibilities as they affect the planning of an air route is of variable importance. If ample facilities are available for instrument navigation, including directive radio ranges or radio compasses, low ceilings and visibilities may be of comparatively little importance except at terminal points. If contact flying must be carried on, however, the ceiling and visibility conditions over the entire route may be of great importance. Conversely, it may be stated that if flying is to be carried out under conditions of poor ceilings and visibilities, ample navigational equipment must be available.

In tropical regions both convectional thunderstorms and hurricanes may adversely affect flying. Fog and low clouds are relatively very rare in the lower latitudes. For daytime flying, convectional thunderstorms may generally be avoided, by slight alteration of the flight path. With night flying, however, this type of storm may cause serious difficulties. If night flying is contemplated, a study of the distribution of convective storms is of considerable value, since it is generally found that they are much more frequent in certain restricted localities than in others. As a general rule, this type of storm is relatively much more frequent to the windward of mountain ranges than to the leeward, and in general the intensity of convective activity is the strongest in regions fairly close to the ocean where abundant moisture is present.

A rather complete meteorological service is essential in tropical regions during the hurricane season. This service should include as many weather reporting stations as possible, and a central forecasting office. Since the early recognition of hurricane centers and the accurate determination of their paths in the early stages is essential, a complete network of weather reporting stations is required. With such a meteorological network, hurricanes should cause little concern to airline operators, since their forecasting has been developed to a fairly satisfactory stage.

CLIMATIC ANALYSES

From the standpoint of aviation, the most satisfactory climatic analyses are those based on *air mass frequencies* over a given region. This type of information is only available for a few localities at the

present time, but will undoubtedly be available for much of the northern hemisphere within the next few years as the use of air mass principles spreads. From this type of climatic chart, regions of maximum interaction between air masses may be found. Areas which are frequently occupied by unstable air, with accompanying thunderstorms and squally weather, will be clearly indicated. The relative frequency of maritime and continental air mass types over any region will be made evident. This is particularly important along coastal areas, where fogginess has a very direct bearing on the location of airways.

Maps showing regions of frontogenesis are also very useful in determining where extratropical storms originate and migrate. These are to be used in close connection with maps of cyclone tracks. It should be noted in this regard that the deformation field alone will not yield the position of the lines of frontogenesis (see page 211).

Thunderstorms—Since thunderstorms have an important bearing on scheduled air transport, in that they affect both the comfort of passengers and the safety of flight, it is of interest to note those portions of a region which are particularly subject to thunderstorm activity. One of the most complete studies of this subject is that which has been carried on by W. H. Alexander for the United States. He has plotted for each month of the year the number of thunderstorms which occurred at various points within the United States. He was thus able to draw isolines for the number of thunderstorm days for a long enough period—1904 to 1933—to establish reliable average values.

Alexander showed that for the year as a whole the maximum occurrence of thunderstorms was centered over central Florida, with another maximum over northern New Mexico. During various months throughout the year the centers of maximum thunderstorm activity shifted to other points. During the month of November, for instance, a pronounced center of activity is found to be located over Arkansas with comparatively few occurring over Florida at this time. During July a great number occur in the southern Rocky Mountain region with a large number also over Florida. In April the center of maximum activity is found to be in the Gulf coast region of Louisiana. The largest number of thunderstorms occur during the month of July with the least number during January. It is interesting to note from Alexander's studies that no part of the

United States is entirely free from thunderstorms. The smallest number occurred at San Francisco with 61 storms for the thirty year period. The greatest number occurred at Tampa where 2820 occurred.

Thunderstorm activity is obviously closely related to the presence of unstable air. It thus attains a maximum in regions subject to frequent invasions of Tropical Gulf air. The high incidence of thunderstorm activity in the southern Rocky Mountain region is explained by the frequent presence of TG air brought in by southeasterly winds over this area during the summer. In portions of the country which are seldom visited by tropical maritime air, the thunderstorm incidence is comparatively low. Thus the northern New England region, and the Pacific coast are visited but rarely by thunderstorms. Furthermore, the general activity is much more pronounced during the summer when unstable TG air masses frequently occupy the entire Gulf coast and Mississippi Valley areas.

SELECTION OF AIRPORT SITES

It is unfortunate that many, if not all, airports have their locations determined by political rather than aeronautical considerations. Since an airport represents a very large investment, it is of the greatest importance that every effort be made to select the most favorable site. Meteorological conditions are of first importance in such a selection. Other matters including,

1. Obstructions, involving the suitability for radio approach systems,
2. Accessibility from the metropolitan area to be served,
3. Cost of the land,

must be given consideration, of course. However, once the location of the most favorable site from the meteorological standpoint is found, the other factors may readily be taken into account.

A thoroughly impartial survey conducted by a board of expert meteorologists, and supported by ample meteorological data is essential to choosing the best site. Once the board's findings are available, they should be given the widest publicity so that pressure by political groups may be nullified. When it is realized that a modern airport represents an investment in land alone of from one hundred

thousand to over a million dollars, and that the total investment in a complete airport may amount to several million dollars, it is readily seen why the strongest political pressure may be brought to bear on its selection. The amount of money spent in obtaining thoroughly reliable data and the services of recognized experts is such a small percentage of the total expenditure that it is a very wise investment.

A survey to determine the most favorable meteorological conditions in a given area requires from one to two years. As a preliminary to establishing observation points, to determine the most favorable weather conditions within a given small area, a study should first be made of: (a) prevailing surface winds, (b) sources of atmospheric pollution, (c) topography, (d) location of water bodies. From a consideration of these factors, the meteorologist can determine which portions of the area are likely to enjoy the most favorable weather. Actual observations will then check this conclusion and furnish additional detailed information.

During at least one year, and preferably for two years, the ceiling, visibility, and general weather conditions should be reported several times per day at the more promising sites. From these observations, it will be possible to determine without question where the most favorable weather conditions exist. In the vicinity of a large city at least ten or fifteen such observation points should be employed. Observations should be made at all airports which are already established in order to forestall any charges of favoritism. Even airports which are quite obviously located in unfavorable meteorological locations may exert much pressure unless facts are available to prove that their weather conditions are poor.

The analysis of the data obtained from the several observation points may be carried out as follows:

1. Weather limitations, below which transport airplanes cannot operate in the general area, must be determined.
2. For each proposed site the total number of hours during which operations would be impossible using these limitations, may then be calculated.

This gives a direct measure of the probable efficiencies of the several sites.

After the most suitable sites from the standpoint of weather conditions are determined, they may then be examined from other

viewpoints by aeronautical experts. It may be found that some compromise with weather conditions will be desirable in order to obtain the highest general efficiency at the airport. Thus, a certain site may enjoy slightly less favorable weather conditions than another, yet be located in an area with superior approaches.

Fog, smoke, and low stratus clouds are the principal weather conditions affecting the choice of an airport site. These are the weather elements which show considerable local variation, and which, therefore, may give rise to widely varying weather conditions within a small area. Ordinary cyclonic storms and thunderstorms show comparatively little average local variation in intensity and therefore are of little importance in the choice of an airport. In mountainous terrain, this is not strictly true, though, since the clouds and precipitation accompanying frontal activity are invariably heavier over higher elevations.

APPENDIX

1. Definitions

Absolute Humidity—Mass of water present in a unit volume of air. Usually measured in grams per cubic meter.

Absolute Instability—A body of air in which the temperature lapse-rate is steeper than either the moist or the dry adiabatic is said to be "absolutely unstable." A particle of air which is given an initial impetus, either upward or downward, within such a body of air, will continue its motion with no further external force necessary.

Absolute Temperature—A temperature scale in which zero lies at Absolute Zero, located at $-273°$ C.

Adiabat—A line on a pressure-temperature diagram which represents adiabatic conditions.

Adiabatic—A physical process which involves no change in heat content.

Air Mass—A body of air with more or less uniform properties throughout in a horizontal direction.

Albedo—Diffuse reflecting power. Ratio of total reflected luminous flux to that received.

Anallobar—A region of maximum pressure rises. An isallobaric HIGH.

Ångstrom Unit—A unit of length equal to 10^{-10} meter. (Usually designated as Å.)

Anticyclone—A type of atmospheric circulation which is characterized by relatively high pressure at the center. The winds blow out of, and around the center. Clockwise in the northern hemisphere, and counterclockwise in the southern hemisphere.

Aphelion—That point in a planet's orbit which is farthest away from the sun. For the earth this occurs on July 1.

Ascendant—A vector which measures the direction and magnitude of the greatest rate of increase of a function.

Austausch—Measure of interchange of heat, momentum, moisture, etc., caused by turbulence.

***Back (wind)*—A counterclockwise change in the wind direction.**

Baroclinic—A fluid in which the surfaces of equal pressure (isobars), and equal density (isosteres) intersect. The quadrilaterals formed by the intersecting surfaces are termed *solenoids*.

Barotropic—A fluid in which the surfaces of equal pressure (isobars), and equal density (isosteres) are parallel. This is the fluid of "classical" hydrodynamics.

Black Body—A perfect radiator. A body which radiates all wave lengths with equal effectiveness. Also, a perfect absorber.

Boyle's Law—At a constant temperature, the volume of a given quantity of any gas varies inversely as the pressure to which the gas is subjected. (P.V. = constant.)

Brickfelder—A Sirocco type wind of Australia.

Charles' Law—At a constant pressure, the volume of a given quantity of any gas varies directly as the temperature to which the gas is subjected. (V/T = constant.)

Chinook—A warm, dynamically heated wind of the Prairie States of the U. S. Same as a foehn wind.

Colloid—A heteorogeneous, two-phase system in which one phase, divided into very small separate volumes, is known as the disperse phase, and the other as the dispersion medium. In meteorology, clouds may be considered as colloids, with the water or ice crystals acting as the disperse phase, and the air as the dispersion medium.

Conditional Instability—A body of air in which the temperature lapse-rate lies between the dry and the moist adiabatic is said to be "conditionally unstable." A particle of air which is lifted in such a body of air will follow a dry adiabat until it becomes saturated, then it will follow a saturated adiabat. Eventually the particle will become warmer (less dense) than its surroundings, and will rise by itself.

Convective Instability—Any body of air, which when lifted, will become eventually absolutely unstable. The air mass curve, when plotted on a Rossby Diagram, has a slope intermediate between the equivalent potential temperature and the partial potential temperature lines.

Coriolis Force—The deviating force on a particle in motion due to the rotation of the earth. Directed to the right in the northern hemisphere, and to the left in the southern hemisphere. Zero at the equator.

Cyclone—A type of atmospheric circulation which is characterized by relatively low pressure at the center. The winds blow into,

and around the center. Counterclockwise in the northern hemisphere, and clockwise in the southern hemisphere.

Dalton's Law—The pressure of a mixture of several gases in a given space is equal to the sum of the partial pressures which each gas would exert if it were confined alone in the space.

Density—A measure of the concentration of matter. Expressed as the mass per unit volume, as grams per cubic centimeter.

Deviating Force—Same as the Coriolis force.

Dew Point—The temperature at which condensation will occur in a given sample of air as it is cooled.

Diurnal—Daily.

Dry Adiabat—A line on a pressure-temperature diagram which represents the rate of cooling of an air particle as it is raised in the atmosphere, *before* it becomes saturated.

Dry Stage—The stage during dynamic cooling of a mass of air, when the temperature remains above the condensation point.

Dynamic Meter—A unit of geopotential equal to 10^5 times the C.G.S. unit of geopotential. It is approximately equal to the unit of length meter. The exact conversion of dynamic meters to meters involves the latitude and elevation above the surface.

Eddy—Any disturbance to the smooth flow of air. May be circular rotating currents, convection currents, or merely irregular disturbances.

Emagram—A thermodynamical diagram with coordinates of temperature and log pressure. Areas are proportional to the energy available.

Entropy—A fundamental concept of thermodynamics which is equal to $\int \frac{dQ}{T}$. It is usually designated by the Greek letter ϕ. It is dependent on the quantity of heat in a body and its temperature. It depends only on the state of the substance, and is independent of the sequence of changes by which that state was reached.

Equinox—The points in the earth's orbit at which the sun crosses the equator. The vernal equinox occurs on March 21, and the autumnal equinox on September 23, in the northern hemisphere. Day and night are equal on these dates, hence the name.

Equivalent Temperature—The temperature a particle of air will assume if lifted until all moisture is precipitated, then lowered to its original level and pressure.

Equivalent Potential Temperature—The temperature a particle of air will assume if lifted until all moisture is precipitated, then lowered to a standard pressure (usually 1000 millibars). Usually stated in Absolute Temperature. (Usually designated as θ_e.)

Foehn Wind—A dynamically heated wind. One whose temperature is raised as the air descends, being warmed according to the general gas law ($P \cdot V = N \cdot R \cdot T$).

Front—A discontinuity surface separating two dissimilar air masses.

Geopotential—The potential energy of unit mass at a point z above the surface of the earth. It is equal to the work done in lifting unit mass from mean sea level up to that point. It is equal to $\int^z gdz$, or approximately, neglecting the variation of g with height, to gz.

Geostrophic Wind—The wind, blowing along the isobars, which will produce a Coriolis force that will just balance the existing pressure gradient. Neglects centrifugal force. See *Gradient Wind.*

Gradient—A vector which measures the direction and magnitude of the greatest rate of decrease of a function.

Gradient Wind—The wind, blowing along the isobars, which will produce a Coriolis force that will just balance the existing pressure gradient. Includes centrifugal force for the general case of curved isobars. The Geostrophic Wind may be regarded as a first approximation to the Gradient Wind.

Grey Body—A body which radiates or absorbs selectively. It radiates (or absorbs) more effectively at one wave length than at others.

Hail Stage—The stage during dynamic cooling of a mass of air, when the temperature remains at the freezing point during the freezing of any liquid water that may be present.

Heaviside Layer—An ionized layer in the upper atmosphere which is important in the propagation of radio signals.

Hertz Diagram—A form of adiabatic chart in which the coordinates are temperature on a vertical linear scale, and pressure on a horizontal log scale.

Hygrometer—An instrument for measuring the humidity.

Ionosphere—The very high levels of the atmosphere, in which the constituent gases are more or less ionized, due to the low pressure. Above about 60 miles. Above the stratosphere.

Isallobar—A line connecting points having equal pressure changes.

Isobar—A line connecting points having equal pressures.

Isopleth—A line connecting points which have the same magnitude of a given function. Thus, an isobar is an isopleth of pressure.

Isostere—A line connecting points having the same density.

Isotherm—A line connecting points having the same temperature.

Isothermal—A physical process which takes place without change in temperature. Opposed to adiabatic, where the process takes place without change in heat content.

Katallobar—A region of maximum pressure falls. An isallobaric LOW.

Kinematics—The branch of mechanics which deals with motions, considered by themselves, without regard to the forces producing them.

Kinetics—The branch of mechanics which deals with the changes of motions which are produced by forces.

Kirchhoff's Law—At a given temperature the ratio between the absorptive and emissive power for a given wave length is the same for all bodies.

Lapse-rate—The rate of decrease of temperature with elevation.

Latent Heat of Fusion—The quantity of heat necessary to change one gram of a solid to a liquid with no change in temperature. Measured in calories per gram.

Latent Heat of Vaporization—The quantity of heat necessary to change one gram of a liquid to vapor with no change in temperature. Measured in calories per gram.

Mechanical Instability—A condition of the atmosphere in which the density increases aloft.

Micron—A unit of length equal to 10^{-6} meter. Usually designated as μ.)

Mistral—A cold cyclonic wind of Europe similar to the *norther* of the United States.

Mixing Ratio—A measure of humidity. The mass of water contained in a given mass of *dry* air. Generally expressed in grams of water per kilogram of dry air. (Usually designated as W.)

Moist Adiabat—A line on a pressure-temperature diagram which

represents the rate of cooling of an air particle as it is raised in the atmosphere, *after* it becomes saturated.

Monsoon—A closed circulation established between the water and adjoining land. Especially, the Indian Monsoon.

Neuhoff Diagram—An adiabatic chart in which the coordinates are temperature on a horizontal linear scale, and pressure on a vertical log scale.

Ozone Layer—A rather restricted region in the upper atmosphere in which most of the ozone is concentrated. The average quantity of ozone in the atmosphere is equivalent to a layer about 3 millimeters thick at standard pressure. Elevation about 25 km.

Pampero—A cold cyclonic wind of South America similar to the *norther* of the United States.

Partial Potential Temperature of the Dry Air—The temperature which a sample of air would assume if lowered (or raised) to a standard pressure of 1000 millibars, neglecting the presence of water vapor. (Usually designated as θ_d.)

Perihelion—That point in a planet's orbit which is closest to the sun. For the earth this occurs on January 1.

Potential Temperature—The temperature which a sample of air would assume if lowered (or raised) to a standard pressure of 1000 millibars. (Usually designated as θ.)

Psychrometer—An instrument for measuring humidity. It consists of two thermometers, the bulb of one of which is kept moistened. When ventilated by forcing air over the thermometer bulbs, it yields two temperature readings—the dry-bulb, and the wet-bulb reading.

Radiosonde—Same as radiometeorograph.

Rain Stage—The stage during dynamic cooling of a mass of air in which liquid water is precipitated.

Refsdal Diagram—An adiabatic chart in which the coordinates are temperature on a horizontal linear scale, and pressure on a vertical log scale.

Relative Humidity—The ratio of the actual vapor pressure to the saturated vapor pressure. (Usually designated as *R.H* or f).

Reynolds Number—A non-dimensional ratio which indicates when turbulence may be expected during fluid flow. It is a measure of the ratio between frictional and dynamic forces. It is equal to $\frac{vl}{\nu}$.

Rossby Diagram—A thermodynamical diagram with coordinates of, specific humidity on a horizontal linear scale, and partial potential temperature of the dry air on a vertical log scale. Sloping lines of constant equivalent potential temperature are also included.

Scalar—A magnitude which has no direction. Time, length, are examples of scalar quantities. See *Vector*.

Scattering—The absorption and re-radiation of incident light by very small particles, or gas molecules. Scattering is very much greater for blue light than for red. The scattering particles must be smaller than the wave length of the light which they are to scatter.

Sirocco—A hot wind blowing in the warm sector of a cyclone; may be either dry or moist, depending on type of air and its trajectory. Type region is Italy.

Snow Stage—The stage during dynamic cooling of a mass of air during which solid water (snow) is precipitated.

Solenoid—The quadrilaterals formed by the intersection of surfaces of equal pressure and equal density.

Solstices—The points in the earth's orbit at which the sun is farthest from the equator. In the northern hemisphere the summer solstice (longest day) occurs about June 21, the winter solstice (shortest day) about December 22. The sun appears to stand still in its course on these dates, hence the name.

Specific Heat—The ratio of the heat capacity of a substance to that of water at 15° C.

Specific Humidity—A measure of humidity. The mass of water contained in a given mass of *moist* air. Generally expressed in grams of water per kilogram of moist air. (Usually designated as Q.)

Stability—The resistance of a body of air to displacements of air particles within it.

Statics—The branch of mechanics which deals with the action of forces on bodies at rest.

Stefan's Law—The amount of energy radiated per unit time from unit surface of a black body is directly proportional to the fourth power of the absolute temperature.

Stratosphere—The upper portion of the atmosphere, above approximately 40,000 to 60,000 feet, where the temperature does not decrease with elevation, and where turbulence is absent. Below the Ionosphere.

Subsidence—The sinking and spreading out of a body of air, usually within an anticyclone.

Tendency—The rate of change of the barometric pressure.

Tephigram—A thermodynamical diagram in which the coordinates are absolute temperature on a horizontal axis and entropy on a vertical axis, both on linear scales. The entropy scale is usually replaced by a log scale of potential temperature, to which it is equivalent.

Tropopause—The boundary between the troposphere and the stratosphere.

Troposphere—The lower portion of the atmosphere between the surface of the earth and the tropopause. Marked by a steady decrease of temperature with elevation, and considerable turbulent mixing.

Typhoon—A tropical cyclonic storm of great intensity. Same as a hurricane.

Vector—A magnitude which also has a direction. Force, acceleration, velocity, are vector quantities. See *Scalar*.

Veer (wind)—A clockwise change in the wind direction.

Virtual Temperature—The temperature of damp air at which *dry air* of the same pressure, would have the same density as the damp air. Always higher than the actual temperature.

Wet-bulb Potential Temperature—The temperature which a sample of air would assume if lowered (or raised) along the moist adiabat through its wet-bulb temperature, to a standard pressure of 1000 millibars.

Wet-bulb Temperature—The temperature indicated by the wet-bulb of a psychrometer.

Zonda—A Sirocco type wind of the Argentine pampas.

APPENDIX

2. Tables

Table 5
Centigrade to Fahrenheit Temperature

Centigrade	0	1	2	3	4	5	6	7	8	9
	F.	F.	F.	F.	F.	F.	F.	F.	F.	F
−50	− 58.0	−59.8	−61.6	− 63.4	−65.2	−67.0	−68.8	−70.6	−72.4	−74.2
−40	− 40.0	41.8	43.6	45.4	47.2	49.0	50.8	52.6	54.4	56.2
−30	− 22.0	23.8	25.6	27.4	29.2	31.0	32.8	34.6	36.4	38.2
−20	− 4.0	− 5.8	− 7.6	− 9.4	−11.2	−13.0	−14.8	−16.6	−18.4	−20.2
−10	+ 14.0	+12.2	+10.4	+ 8.6	+ 6.8	+ 5.0	+ 3.2	+ 1.4	− 0.4	− 2.2
− 0	+ 32.0	+30.2	+28.4	+26.6	+24.8	+23.0	+21.2	+19.4	+17.6	+15.8
+ 0	+ 32.0	+33.8	+35.6	+37.4	+39.2	+41.0	+42.8	+44.6	+46.4	+48.2
+10	+ 50.0	51.8	53.6	55.4	57.2	59.0	60.8	62.6	64.4	66.2
+20	+ 68.0	69.8	71.6	73.4	75.2	77.0	78.8	80.6	82.4	84.2
+30	+ 86.0	87.8	89.6	91.4	93.2	95.0	96.8	98.6	100.4	102.2
+40	+104.0	105.8	107.6	109.4	111.2	113.0	114.8	116.6	118.4	120.2
+50	+122.0	123.8	125.6	127.4	129.2	131.0	132.8	134.6	136.4	138.2

Table 6
Fahrenheit to Centigrade Temperature

Fahrenheit	0	1	2	3	4	5	6	7	8	9
	C.	C.	C.	C.	C.	C.	C.	C.	C.	C.
− 70°	−56.7	−57.2	−57.8	−58.3	−58.9	−59.4	−60.0	−60.6	−61.1	−61.7
− 60	−51.1	51.7	52.2	52.8	53.3	53.9	54.4	55.0	55.6	56.1
− 50	−45.6	46.1	46.7	47.2	47.8	48.3	48.9	49.4	50.0	50.6
− 40	−40.0	40.6	41.1	41.7	42.2	42.8	43.3	43.9	44.4	45.0
− 30	−34.4	35.0	35.6	36.1	36.7	37.2	37.8	38.3	38.9	39.4
− 20	−28.9	29.4	30.0	30.6	31.1	31.7	32.2	32.8	33.3	33.9
− 10	−23.3	23.9	24.4	25.0	25.6	26.1	26.7	27.2	27.8	28.3
− 0	−17.8	18.3	18.9	19.4	20.0	20.6	21.1	21.7	22.2	22.8
+ 0	−17.8	−17.2	−16.7	−16.1	−15.6	−15.0	−14.4	−13.9	−13.3	−12.8
+ 10	−12.2	11.7	11.1	10.6	10.0	9.4	8.9	8.3	7.8	7.2
+ 20	− 6.7	6.1	5.6	5.0	4.4	3.9	3.3	2.8	2.2	1.7
+ 30	− 1.1	− 0.6	0.0	+ 0.6	+ 1.1	+ 1.7	+ 2.2	+ 2.8	+ 3.3	+ 3.9
+ 40	+ 4.4	+ 5.0	+ 5.6	+ 6.1	+ 6.7	+ 7.2	+ 7.8	+ 8.3	+ 8.9	+ 9.4
+ 50	+10.0	10.6	11.1	11.7	12.2	12.8	13.3	13.9	14.4	15.0
+ 60	+15.6	16.1	16.7	17.2	17.8	18.3	18.9	19.4	20.0	20.6
+ 70	21.1	21.7	22.2	22.8	23.3	23.9	24.4	25.0	25.6	26.1
+ 80	26.7	27.2	27.8	28.3	28.9	29.4	30.0	30.6	31.1	31.7
+ 90	32.2	32.8	33.3	33.9	34.4	35.0	35.6	36.1	36.7	37.2
+100	37.8	38.3	38.9	39.4	40.0	40.6	41.1	41.7	42.2	42.8
+110	43.3	43.9	44.4	45.0	45.6	46.1	46.7	47.2	47.8	48.3
+120	48.9	49.4	50.0	50.6	51.1	51.7	52.2	52.8	53.3	53.9

Table 7

Millibars to Inches of Mercury

1 mb. = 0.0295299 inch.

Millibars	0	1	2	3	4	5	6	7	8	9
	Inches	Inches	Inches	Inches	Inches	Inches	Inches	Inches	Inches	Inches
930	27.46	27.49	27.52	27.55	27.58	27.61	27.64	27.67	27.70	27.73
940	27.76	27.79	27.82	27.85	27.88	27.91	27.94	27.97	28.00	28.03
950	28.05	28.08	28.11	28.14	28.17	28.20	28.23	28.26	28.29	28.32
960	28.35	28.38	28.41	28.44	28.47	28.50	28.53	28.56	28.59	28.62
970	28.65	28.67	28.70	28.73	28.76	28.79	28.82	28.85	28.88	28.91
980	28.94	28.97	29.00	29.03	29.06	29.09	29.12	29.15	29.18	29.21
990	29.24	29.26	29.29	29.32	29.35	29.38	29.41	29.44	29.47	29.50
1000	29.53	29.56	29.59	29.62	29.65	29.68	29.71	29.74	29.77	29.80
1010	29.83	29.86	29.89	29.92	29.94	29.97	30.00	30.03	30.06	30.09
1020	30.12	30.15	30.18	30.21	30.24	30.27	30.30	30.33	30.36	30.39
1030	30.42	30.45	30.48	30.51	30.53	30.56	30.59	30.62	30.65	30.68
1040	30.71	30.74	30.77	30.80	30.83	30.86	30.89	30.92	30.95	30.98
1050	31.01	31.04	31.07	31.10	31.13	31.15	31.18	31.21	31.24	31.27
1060	31.30	31.33	31.36	31.39	31.42	31.45	31.48	31.51	31.54	31 57

Table 8

Miscellaneous Constants and Units

Pressure

Standard atmospheric pressure	= 1,013,250.144 dynes/cm^2
	= 1,013.250144 mb
	= 760 mm Hg at 0° C. and standard gravity
	= 29.92117 in. Hg at 0° C. and standard gravity
	= 14.69603 lb/in^2

Gravity

Standard acceleration due to gravity	= 980.665 cm/sec^2
	= 32.1740 ft/sec^2

Density

Density of dry air with normal content of CO_2 (3 volumes CO_2 per 10,000 of air) at 760 mm pressure and 0° C. temperature	= 0.0012930 g/cm^3
Density of dry air, free from CO_2, at 760 mm pressure and 0° C. temperature	= 0.0012928 g/cm^3
Density of Mercury (Hg) at 0° C.	= 13.595 g/cm^3

Heat

Specific heat of dry air at constant pressure (c_p)	= 0.2396 g-cal/g
Specific heat of dry air at constant volume (c_v)	= 0.1707 g-cal/g
Ratio of specific heats, γ (c_p/c_v)	= 1.403
Latent heat of fusion of ice	= 79.7 g-cal/g
Latent heat of vaporization of water (See Table 19)	
Universal gas constant (R)	= 8.317 × 10^7 ergs/°C./g mole
	= 8.317 Joules/°C./g mole
Specific gas constant for dry air:	
Units: Millibars, kg/m^3, °A.(C.)	= 2.8703
Dynes/cm^2, gm/cm^3, °A.(C.)	= 2.8703 × 10^6
cm of Hg, kg/m^3, °A.(C.)	= 0.215
in of Hg, lb/ft^3, °A. (F.)	= 0.751

Conversions

1 meter	= 39.3700 in
	= 3.280833 ft
1 kilogram	= 2.204622 lb
1 mm Hg 0° C. and standard gravity	= 1.333224 mb
1 inch Hg, 0° C. and standard gravity	= 33.863953 mb

Table 9

Millimeters into Inches

1 mm. = 0.03937 inch.

Millimeters.	0	1	2	3	4	5	6	7	8	9
	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.
0	0.0000	0.0394	0.0787	0.1181	0.1575	0.1968	0.2362	0.2756	0.3150	0.3543
10	0.3937	0.4331	0.4724	0.5118	0.5512	0.5906	0.6299	0.6693	0.7087	0.7480
20	0.7874	0.8268	0.8661	0.9055	0.9449	0.9842	1.0236	1.0630	1.1024	1.1417
30	1.1811	1.2205	1.2598	1.2992	1.3386	1.3780	1.4173	1.4567	1.4961	1.5354
40	1.5748	1.6142	1.6535	1.6929	1.7323	1.7716	1.8110	1.8504	1.8898	1.9291
50	1.9685	2.0079	2.0472	2.0866	2.1260	2.1654	2.2047	2.2441	2.2835	2.3228
60	2.3622	2.4016	2.4409	2.4803	2.5197	2.5590	2.5984	2.6378	2.6772	2.7165
70	2.7559	2.7953	2.8346	2.8740	2.9134	2.9528	2.9921	3.0315	3.0709	3.1102
80	3.1496	3.1890	3.2283	3.2677	3.3071	3.3464	3.3858	3.4252	3.4646	3.5039
90	3.5433	3.5828	3.6220	3.6614	3.7008	3.7402	3.7795	3.8189	3.8583	3.8976
710	27.953	27.957	27.961	27.965	27.968	27.972	27.976	27.980	27.984	27.988
711	27.992	27.996	28.000	28.004	28.008	28.012	28.016	28.020	28.024	28.028
712	28.031	28.035	28.039	28.043	28.047	28.051	28.055	28.059	28.063	28.067
713	28.071	28.075	28 079	28.083	28.087	28.090	28.094	28.098	28.102	28.106
714	28.110	28.114	28.118	28.122	28.126	28.130	28.134	28.138	28.142	28.146
715	28.150	28.153	28.157	28.161	28.165	28.169	28.173	28.177	28.181	28.185
716	28.189	28.193	28.197	28.201	28.205	28.209	28.213	28.216	28.220	28.224
717	28.228	28.232	28.236	28.240	28.244	28.248	28.252	28.256	28.260	28.264
718	28.268	28.272	28.276	28.279	28.283	28.287	28.291	28.295	28.299	28.303
719	28.307	28.311	28.315	28.319	28.323	28.327	28.331	28.335	28.339	28.342
720	28.346	28.350	28.354	28.358	28.362	28.366	28.370	28.374	28.378	28.382
721	28.386	28.390	28.394	28.398	28.402	28.405	28.409	28.413	28.417	28.421
722	28.425	28.429	28.433	28.437	28.441	28.445	28.449	28.453	28.457	28.461
723	28.465	28.468	28.472	28.476	28.480	28.484	28.488	28.492	28.496	28.500
724	28.504	28.508	28.512	28.516	28.520	28.524	28.528	28.531	28.535	28.539
725	28.543	28.547	28.551	28.555	28.559	28.563	28.567	28.571	28.575	28.579
726	28.583	28.587	28.590	28.594	28.598	28.602	28.606	28.610	28.614	28.618
727	28.622	28.626	28.630	28.634	28.638	28.642	28.646	28.650	28.653	28.657
728	28.661	28.665	28.669	28.673	28.677	28.681	28.685	28.689	28.693	28.697
729	28.701	28.705	28.709	28.713	28.716	28.720	28.724	28.728	28.732	28.736
730	28.740	28.744	28.748	28.752	28.756	28.760	28.764	28.768	28.772	28.776
731	28.779	28.783	28.787	28.791	28.795	28.799	28.803	28.807	28.811	28.815
732	28.819	28.823	28.827	28.831	28.835	28.839	28.842	28.846	28.850	28.854
733	28.858	28.862	28.866	28.870	28.874	28.878	28.882	28.886	28.890	28.894
734	28.898	28.902	28.905	28.909	28.913	28.917	28.921	28.925	28.929	28.933
735	28.937	28.941	28.945	28.949	28.953	28.957	28.961	28.965	28.968	28.972
736	28.976	28.980	28.984	28.988	28.992	28.996	29.000	29.004	29.008	29.012
737	29.016	29.020	29.024	29.028	29.031	29.035	29.039	29.043	29.047	29.051
738	29.055	29.059	29.063	29.067	29.071	29.075	29.079	29.083	29.087	29.090
739	29.094	29.098	29.102	29.106	29.110	29.114	29.118	29.122	29.126	29.130
740	29.134	29.138	29.142	29.146	29.150	29.153	29.157	29.161	29.165	29.169
741	29.173	29.177	29.181	29.185	29.189	29.193	29.197	29.201	29.205	29.209
742	29.213	29.216	29.220	29.224	29.228	29.232	29.236	29.240	29.244	29.248
743	29.252	29.256	29.260	29.264	29.268	29.272	29.276	29.279	29.283	29.287
744	29.291	29.295	29.299	29.303	29.307	29.311	29.315	29.319	29.323	29.327
745	29.331	29.335	29.339	29.342	29.346	29.350	29.354	29.358	29.362	29.366
746	29.370	29.374	29.378	29.382	29.386	29.390	29.394	29.398	29.402	29.405
747	29.409	29.413	29.417	29.421	29.425	29.429	29.433	29.437	29.441	29.445
748	29.449	29.453	29.457	29.461	29.465	29.468	29.472	29.476	29.480	29.484
749	29.488	29.492	29.496	29.500	29.504	29.508	29.512	29.516	29.520	29.524
750	29.528	29.531	29.535	29.539	29.543	29.547	29.551	29.555	29.559	29.563

TABLE 9—*Continued*

MILLIMETERS INTO INCHES—*Continued*

1 mm. = 0.03937 inch.

Milli-meters.	.0	.1	.2	.3	.4	.5	.6	.7	.8	.9
	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.	Inches.
750	29.528	29.531	29.535	29.539	29.543	29.547	29.551	29.555	29.559	29.563
751	29.567	29.571	29.575	29.579	29.583	29.587	29.590	29.594	29.598	29.602
752	29.606	29.610	29.614	29.618	29.622	29.626	29.630	29.634	29.638	29.642
753	29.646	29.650	29.653	29.657	29.661	29.665	29.669	29.673	29.677	29.681
754	29.685	29.689	29.693	29.697	29.701	29.705	29.709	29.713	29.716	29.720
755	29.724	29.728	29.732	29.736	29.740	29.744	29.748	29.752	29.756	29.760
756	29.764	29.768	29.772	29.776	29.779	29.783	29.787	29.791	29.795	29.799
757	29.803	29.807	29.811	29.815	29.819	29.823	29.827	29.831	29.835	29.839
758	29.842	29.846	29.850	29.854	29.858	29.862	29.866	29.870	29.874	29.878
759	29.882	29.886	29.890	29.894	29.898	29.902	29.905	29.909	29.913	29.917
760	29.921	29.925	29.929	29.933	29.937	29.941	29.945	29.949	29.953	29.957
761	29.961	29.965	29.968	29.972	29.976	29.980	29.984	29.988	29.992	29.996
762	30.000	30.004	30.008	30.012	30.016	30.020	30.024	30.027	30.031	30.035
763	30.039	30.043	30.047	30.051	30.055	30.059	30.063	30.067	30.071	30.075
764	30.079	30.083	30.087	30.090	30.094	30.098	30.102	30.106	30.110	30.114
765	30.118	30.122	30.126	30.130	30.134	30.138	30.142	30.146	30.150	30.153
766	30.157	30.161	30.165	30.169	30.173	30.177	30.181	30.185	30.189	30.193
767	30.197	30.201	30.205	30.209	30.213	30.216	30.220	30.224	30.228	30.232
768	30.236	30.240	30.244	30.248	30.252	30.256	30.260	30.264	30.268	30.272
769	30.276	30.279	30.283	30.287	30.291	30.295	30.299	30.303	30.307	30.311
770	30.315	30.319	30.323	30.327	30.331	30.335	30.339	30.342	30.346	30.350
771	30.354	30.358	30.362	30.366	30.370	30.374	30.378	30.382	30.386	30.390
772	30.394	30.398	30.402	30.405	30.409	30.413	30.417	30.421	30.425	30.429
773	30.433	30.437	30.441	30.445	30.449	30.453	30.457	30.461	30.465	30.468
774	30.472	30.476	30.480	30.484	30.488	30.492	30.496	30.500	30.504	30.508
775	30.512	30.516	30.520	30.524	30.528	30.531	30.535	30.539	30.543	30.547
776	30.551	30.555	30.559	30.563	30.567	30.571	30.575	30.579	30.583	30.587
777	30.590	30.594	30.598	30.602	30.606	30.610	30.614	30.618	30.622	30.626
778	30.630	30.634	30.638	30.642	30.646	30.650	30.653	30.657	30.661	30.665
779	30.669	30.673	30.677	30.681	30.685	30.689	30.693	30.697	30.701	30.705
780	30.709	30.713	30.716	30.720	30.724	30.728	30.732	30.736	30.740	30.744
781	30.748	30.752	30.756	30.760	30.764	30.768	30.772	30.776	30.779	30.783
782	30.787	30.791	30.795	30.799	30.803	30.807	30.811	30.815	30.819	30.823
783	30.827	30.831	30.835	30.839	30.842	30.846	30.850	30.854	30.858	30.862
784	30.866	30.870	30.874	30.878	30.882	30.886	30.890	30.894	30.898	30.902
785	30.905	30.909	30.913	30.917	30.921	30.925	30.929	30.933	30.937	30.941
786	30.945	30.949	30.953	30.957	30.961	30.965	30.968	30.972	30.976	30.980
787	30.984	30.988	30.992	30.996	31.000	31.004	31.008	31.012	31.016	31.020
788	31.024	31.027	31.031	31.035	31.039	31.043	31.047	31.051	31.055	31.059
789	31.063	31.067	31.071	31.075	31.079	31.083	31.087	31.090	31.094	31.098
790	31.102	31.106	31.110	31.114	31.118	31.122	31.126	31.130	31.134	31.138
791	31.142	31.146	31.150	31.153	31.157	31.161	31.165	31.169	31.173	31.177
792	31.181	31.185	31.189	31.193	31.197	31.201	31.205	31.209	31.213	31.216
793	31.220	31.224	31.228	31.232	31.236	31.240	31.244	31.248	31.252	31.256
794	31.260	31.264	31.268	31.272	31.276	31.279	31.283	31.287	31.291	31.295
795	31.299	31.303	31.307	31.311	31.315	31.319	31.323	31.327	31.331	31.335
796	31.339	31.342	31.346	31.350	31.354	31.358	31.362	31.366	31.370	31.374
797	31.378	31.382	31.380	31.390	31.394	31.398	31.402	31.405	31.409	31.413
798	31.417	31.421	31.425	31.429	31.433	31.437	31.441	31.445	31.449	31.453
799	31.457	31.461	31.465	31.468	31.472	31.476	31.480	31.484	31.488	31.492
800	31.496	31.500	31.504	31.508	31.512	31.516	31.520	31.524	31.527	31.531

Table 10

Barometric Inches (Mercury) into Millibars

1 inch = 33.86395 mb.

Inches	.00	.01	.02	.03	.04	.05	.06	.07	.08	.09
	mb.	mb.	mb.	mb.	mb.	mb.	mb.	mb.	mb.	mb.
0.0	0.00	0.34	0.68	1.02	1.35	1.69	2.03	2.37	2.71	3.05
0.1	3.39	3.73	4.06	4.40	4.74	5.08	5.42	5.76	6.10	6.43
0.2	6.77	7.11	7.45	7.79	8.13	8.47	8.80	9.14	9.48	9.82
0.3	10.16	10.50	10.84	11.18	11.51	11.85	12.19	12.53	12.87	13.21
0.4	13.55	13.88	14.22	14.56	14.90	15.24	15.58	15.92	16.25	16.59
0.5	16.93	17.27	17.61	17.95	18.29	18.63	18.96	19.30	19.64	19.98
0.6	20.32	20.66	21.00	21.33	21.67	22.01	22.35	22.69	23.03	23.37
0.7	23.70	24.04	24.38	24.72	25.06	25.40	25.74	26.08	26.41	26.75
0.8	27.09	27.43	27.77	28.11	28.45	28.78	29.12	29.46	29.80	30.14
0.9	30.48	30.82	31.15	31.49	31.83	32.17	32.51	32.85	33.19	33.53
1.0	33.86	34.20	34.54	34.88	35.22	35.56	35.90	36.23	36.57	36.91
1.1	37.25	37.59	37.93	38.27	38.60	38.94	39.28	39.62	39.96	40.30
1.2	40.64	40.98	41.31	41.65	41.99	42.33	42.67	43.01	43.35	43.68
1.3	44.02	44.36	44.70	45.04	45.38	45.72	46.05	46.39	46.73	47.07
1.4	47.41	47.75	48.09	48.43	48.76	49.10	49.44	49.78	50.12	50.46
1.5	50.80	51.13	51.47	51.81	52.15	52.49	52.83	53.17	53.51	53.84
1.6	54.18	54.52	54.86	55.20	55.54	55.88	56.21	56.55	56.89	57.23
1.7	57.57	57.91	58.25	58.58	58.92	59.26	59.60	59.94	60.28	60.62
1.8	60.96	61.29	61.63	61.97	62.31	62.65	62.99	63.33	63.66	64.00
1.9	64.34	64.68	65.02	65.36	65.70	66.03	66.37	66.71	67.05	67.39
2.0	67.73	68.07	68.41	68.74	69.08	69.42	69.76	70.10	70.44	70.78
2.1	71.11	71.45	71.79	72.13	72.47	72.81	73.15	73.48	73.82	74.16
2.2	74.50	74.84	75.18	75.52	75.86	76.19	76.53	76.87	77.21	77.55
2.3	77.89	78.23	78.56	78.90	79.24	79.58	79.92	80.26	80.60	80.93
2.4	81.27	81.61	81.95	82.29	82.63	82.97	83.31	83.64	83.98	84.32
25.0	846.6	846.9	847.3	847.6	848.0	848.3	848.6	849.0	849.3	849.6
25.1	850.0	850.3	850.7	851.0	851.3	851.7	852.0	852.4	852.7	853.0
25.2	853.4	853.7	854.0	854.4	854.7	855.1	855.4	855.7	856.1	856.4
25.3	856.8	857.1	857.4	857.8	858.1	858.5	858.8	859.1	859.5	859.8
25.4	860.1	860.5	860.8	861.2	861.5	861.8	862.2	862.5	862.9	863.2
25.5	863.5	863.9	864.2	864.5	864.9	865.2	865.6	865.9	866.2	866.6
25.6	866.9	867.3	867.6	867.9	868.3	868.6	868.9	869.3	869.6	870.0
25.7	870.3	870.7	871.0	871.3	871.7	872.0	872.3	872.7	873.0	873.4
25.8	873.7	874.0	874.4	874.7	875.0	875.4	875.7	876.1	876.4	876.7
25.9	877.1	877.4	877.8	878.1	878.4	878.8	879.1	879.4	879.8	880.1
26.0	880.5	880.8	881.1	881.5	881.8	882.2	882.5	882.8	883.2	883.5
26.1	883.8	884.2	884.5	884.9	885.2	885.5	885.9	886.2	886.6	886.9
26.2	887.2	887.6	887.9	888.3	888.6	888.9	889.3	889.6	889.9	890.3
26.3	890.6	891.0	891.3	891.6	892.0	892.3	892.7	893.0	893.3	893.7
26.4	894.0	894.3	894.7	895.0	895.4	895.7	896.0	896.4	896.7	897.1
26.5	897.4	897.7	898.1	898.4	898.7	899.1	899.4	899.8	900.1	900.4
26.6	900.8	901.1	901.5	901.8	902.1	902.5	902.8	903.2	903.5	903.8
26.7	904.2	904.5	904.8	905.2	905.5	905.9	906.2	906.5	906.9	907.2
26.8	907.6	907.9	908.2	908.6	908.9	909.2	909.6	909.9	910.3	910.6
26.9	910.9	911.3	911.6	912.0	912.3	912.6	913.0	913.3	913.6	914.0
27.0	914.3	914.7	915.0	915.3	915.7	916.0	916.4	916.7	917.0	917.4
27.1	917.7	918.1	918.4	918.7	919.1	919.4	919.7	920.1	920.4	920.8
27.2	921.1	921.4	921.8	922.1	922.5	922.8	923.1	923.5	923.8	924.1
27.3	924.5	924.8	925.2	925.5	925.8	926.2	926.5	926.9	927.2	927.5
27.4	927.9	928.2	928.5	928.9	929.2	929.6	929.9	930.2	930.6	930.9

Table 10—*Continued*

Barometric Inches (Mercury) into Millibars

Inches.	.00	.01	.02	.03	.04	.05	.06	.07	.08	.09
	mb.	mb.	mb.	mb.	mb.	mb.	mb.	mb.	mb.	mb.
27.5	931.3	931.6	931.9	932.3	932.6	933.0	933.3	933.6	934.0	934.3
27.6	934.6	935.0	935.3	935.7	936.0	936.3	936.7	937.0	937.4	937.7
27.7	938.0	938.4	938.7	939.0	939.4	939.7	940.1	940.4	940.7	941.1
27.8	941.4	941.8	942.1	942.4	942.8	943.1	943.4	943.8	944.1	944.5
27.9	944.8	945.1	945.5	945.8	946.2	946.5	946.8	947.2	947.5	947.9
28.0	948.2	948.5	948.9	949.2	949.5	949.9	950.2	950.6	950.9	951.2
28.1	951.6	951.9	952.3	952.6	952.9	953.3	953.6	953.9	954.3	954.6
28.2	955.0	955.3	955.6	956.0	956.3	956.7	957.0	957.3	957.7	958.0
28.3	958.3	958.7	959.0	959.4	959.7	960.0	960.4	960.7	961.1	961.4
28.4	961.7	962.1	962.4	962.8	963.1	963.4	963.8	964.1	964.4	964.8
28.5	965.1	965.5	965.8	966.1	966.5	966.8	967.2	967.5	967.8	968.2
28.6	968.5	968.8	969.2	969.5	969.9	970.2	970.5	970.9	971.2	971.6
28.7	971.9	972.2	972.6	972.9	973.2	973.6	973.9	974.3	974.6	974.9
28.8	975.3	975.6	976.0	976.3	976.6	977.0	977.3	977.7	978.0	978.3
28.9	978.7	979.0	979.3	979.7	980.0	980.4	980.7	981.0	981.4	981.7
29.0	982.1	982.4	982.7	983.1	983.4	983.7	984.1	984.4	984.8	985.1
29.1	985.4	985.8	986.1	986.5	986.8	987.1	987.5	987.8	988.2	988.5
29.2	988.8	989.2	989.5	989.8	990.2	990.5	990.9	991.2	991.5	991.9
29.3	992.2	992.6	992.9	993.2	993.6	993.9	994.2	994.6	994.9	995.3
29.4	995.6	995.9	996.3	996.6	997.0	997.3	997.6	998.0	998.3	998.6
29.5	999.0	999.3	999.7	1000.0	1000.4	1000.7	1001.0	1001.4	1001.7	1002.0
29.6	1002.4	1002.7	1003.1	1003.4	1003.7	1004.1	1004.4	1004.7	1005.1	1005.4
29.7	1005.8	1006.1	1006.4	1006.8	1007.1	1007.5	1007.8	1008.1	1008.5	1008.8
29.8	1009.1	1009.5	1009.8	1010.2	1010.5	1010.8	1011.2	1011.5	1011.9	1012.2
29.9	1012.5	1012.9	1013.2	1013.5	1013.9	1014.2	1014.6	1014.9	1015.2	1015.6
30.0	1015.9	1016.3	1016.6	1016.9	1017.3	1017.6	1018.0	1018.3	1018.6	1019.0
30.1	1019.3	1019.6	1020.0	1020.3	1020.7	1021.0	1021.3	1021.7	1022.0	1022.4
30.2	1022.7	1023.0	1023.4	1023.7	1024.0	1024.4	1024.7	1025.1	1025.4	1025.7
30.3	1026.1	1026.4	1026.8	1027.1	1027.4	1027.8	1028.1	1028.4	1028.8	1029.1
30.4	1029.5	1029.8	1030.1	1030.5	1030.8	1031.2	1031.5	1031.8	1032.2	1032.5
30.5	1032.9	1033.2	1033.5	1033.9	1034.2	1034.5	1034.9	1035.2	1035.6	1035.9
30.6	1036.2	1036.6	1036.9	1037.3	1037.6	1037.9	1038.3	1038.6	1038.9	1039.3
30.7	1039.6	1040.0	1040.3	1040.6	1041.0	1041.3	1041.7	1042.0	1042.3	1042.7
30.8	1043.0	1043.3	1043.7	1044.0	1044.4	1044.7	1045.0	1045.4	1045.7	1046.1
30.9	1046.4	1046.7	1047.1	1047.4	1047.8	1048.1	1048.4	1048.8	1049.1	1049.5
31.0	1049.8	1050.1	1050.5	1050.8	1051.1	1051.5	1051.8	1052.2	1052.5	1052.8
31.1	1053.2	1053.5	1053.8	1054.2	1054.5	1054.9	1055.2	1055.5	1055.9	1056.2
31.2	1056.6	1056.9	1057.2	1057.6	1057.9	1058.2	1058.6	1058.9	1059.3	1059.6
31.3	1059.9	1060.3	1060.6	1061.0	1061.3	1061.6	1062.0	1062.3	1062.7	1063.0
31.4	1063.3	1063.7	1064.0	1064.3	1064.7	1065.0	1065.4	1065.7	1066.0	1066.4
31.5	1066.7	1067.1	1067.4	1067.7	1068.1	1068.4	1068.7	1069.1	1069.4	1069.8
31.6	1070.1	1070.4	1070.8	1071.1	1071.5	1071.8	1072.1	1072.5	1072.8	1073.1
31.7	1073.5	1073.8	1074.2	1074.5	1074.8	1075.2	1075.5	1075.9	1076.2	1076.5
31.8	1076.9	1077.2	1077.6	1077.9	1078.2	1078.6	1078.9	1079.2	1079.6	1079.9
31.9	1080.3	1080.6	1080.9	1081.3	1081.6	1082.0	1082.3	1082.6	1083.0	1083.3

Table II

Barometric Millimeters (Mercury) into Millibars

1 mm. = 1.33322387 mb.

Milli-meters.	0	1	2	3	4	5	6	7	8	9
	mb.	mb.	mb.	mb.	mb.	mb.	mb.	mb.	mb.	mb.
0	0	1.3	2.7	4.0	5.3	6.7	8.0	9.3	10.7	12.0
10	13.3	14.7	16.0	17.3	18.7	20.0	21.3	22.7	24.0	25.3
20	26.7	28.0	29.3	30.7	32.0	33.3	34.7	36.0	37.3	38.7
30	40.0	41.3	42.7	44.0	45.3	46.7	48.0	49.3	50.7	52.0
40	53.3	54.7	56.0	57.3	58.7	60.0	61.3	62.7	64.0	65.3
50	66.7	68.0	69.3	70.7	72.0	73.3	74.7	76.0	77.3	78.7
60	80.0	81.3	82.7	84.0	85.3	86.7	88.0	89.3	90.7	92.0
70	93.3	94.7	96.0	97.3	98.7	100.0	101.3	102.7	104.0	105.3
80	106.7	108.0	109.3	110.7	112.0	113.3	114.7	116.0	117.3	118.7
90	120.0	121.3	122.7	124.0	125.3	126.7	128.0	129.3	130.7	132.0
100	133.3	134.7	136.0	137.3	138.7	140.0	141.3	142.7	144.0	145.3
110	146.7	148.0	149.3	150.7	152.0	153.3	154.7	156.0	157.3	158.7
120	160.0	161.3	162.7	164.0	165.3	166.7	168.0	169.3	170.7	172.0
130	173.3	174.7	176.0	177.3	178.7	180.0	181.3	182.7	184.0	185.3
140	186.7	188.0	189.3	190.7	192.0	193.3	194.7	196.0	197.3	198.7
150	200.0	201.3	202.7	204.0	205.3	206.6	208.0	209.3	210.6	212.0
160	213.3	214.6	216.0	217.3	218.6	220.0	221.3	222.6	224.0	225.3
170	226.6	228.0	229.3	230.6	232.0	233.3	234.6	236.0	237.3	238.6
180	240.0	241.3	242.6	244.0	245.3	246.6	248.0	249.3	250.6	252.0
190	253.3	254.6	256.0	257.3	258.6	260.0	261.3	262.6	264.0	265.3
200	266.6	268.0	269.3	270.6	272.0	273.3	274.6	276.0	277.3	278.6
210	280.0	281.3	282.6	284.0	285.3	286.6	288.0	289.3	290.6	292.0
220	293.3	294.6	296.0	297.3	298.6	300.0	301.3	302.6	304.0	305.3
230	306.6	308.0	309.3	310.6	312.0	313.3	314.6	316.0	317.3	318.6
240	320.0	321.3	322.6	324.0	325.3	326.6	328.0	329.3	330.6	332.0
250	333.3	334.6	336.0	337.3	338.6	340.0	341.3	342.6	344.0	345.3
260	346.6	348.0	349.3	350.6	352.0	353.3	354.6	356.0	357.3	358.6
270	360.0	361.3	362.6	364.0	365.3	366.6	368.0	369.3	370.6	372.0
280	373.3	374.6	376.0	377.3	378.6	380.0	381.3	382.6	384.0	385.3
290	386.6	388.0	389.3	390.6	392.0	393.3	394.6	396.0	397.3	398.6
300	400.0	401.3	402.6	404.0	405.3	406.6	408.0	409.3	410.6	412.0
310	413.3	414.6	416.0	417.3	418.6	420.0	421.3	422.6	424.0	425.3
320	426.6	428.0	429.3	430.6	432.0	433.3	434.6	436.0	437.3	438.6
330	440.0	441.3	442.6	444.0	445.3	446.6	448.0	449.3	450.6	452.0
340	453.3	454.6	456.0	457.3	458.6	460.0	461.3	462.6	464.0	465.3
350	466.6	468.0	469.3	470.6	472.0	473.3	474.6	476.0	477.3	478.6
360	480.0	481.3	482.6	484.0	485.3	486.6	488.0	489.3	490.6	492.0
370	493.3	494.6	496.0	497.3	498.6	500.0	501.3	502.6	504.0	505.3
380	506.6	508.0	509.3	510.6	512.0	513.3	514.6	516.0	517.3	518.6
390	520.0	521.3	522.6	524.0	525.3	526.6	528.0	529.3	530.6	532.0
400	533.3	534.6	536.0	537.3	538.6	540.0	541.3	542.6	544.0	545.3
410	546.6	548.0	549.3	550.6	552.0	553.3	554.6	556.0	557.3	558.6
420	560.0	561.3	562.6	564.0	565.3	566.6	568.0	569.3	570.6	572.0
430	573.3	574.6	576.0	577.3	578.6	580.0	581.3	582.6	584.0	585.3
440	586.6	588.0	589.3	590.6	592.0	593.3	594.6	596.0	597.3	598.6

TABLE 11—*Continued*

BAROMETRIC MILLIMETERS (MERCURY) INTO MILLIBARS

Milli-meters.	0	1	2	3	4	5	6	7	8	9
	mb.	mb.	mb.	mb.	mb.	mb.	mb.	mb.	mb.	mb.
450	600.0	601.3	602.6	604.0	605.3	606.6	608.0	609.3	610.6	611.9
460	613.3	614.6	615.9	617.3	618.6	619.9	621.3	622.6	623.9	625.3
470	626.6	627.9	629.3	630.6	631.9	633.3	634.6	635.9	637.3	638.6
480	639.9	641.3	642.6	643.9	645.3	646.6	647.9	649.3	650.6	651.9
490	653.3	654.6	655.9	657.3	658.6	659.9	661.3	662.6	663.9	665.3
500	666.6	667.9	669.3	670.6	671.9	673.3	674.6	675.9	677.3	678.6
510	679.9	681.3	682.6	683.9	685.3	686.6	687.9	689.3	690.6	691.9
520	693.3	694.6	695.9	697.3	698.6	699.9	701.3	702.6	703.9	705.3
530	706.6	707.9	709.3	710.6	711.9	713.3	714.6	715.9	717.3	718.6
540	719.9	721.3	722.6	723.9	725.3	726.6	727.9	729.3	730.6	731.9
550	733.3	734.6	735.9	737.3	738.6	739.9	741.3	742.6	743.9	745.3
560	746.6	747.9	749.3	750.6	751.9	753.3	754.6	755.9	757.3	758.6
570	759.9	761.3	762.6	763.9	765.3	766.6	767.9	769.3	770.6	771.9
580	773.3	774.6	775.9	777.3	778.6	779.9	781.3	782.6	783.9	785.3
590	786.6	787.9	789.3	790.6	791.9	793.3	794.6	795.9	797.3	798.6
600	799.9	801.3	802.6	803.9	805.3	806.6	807.9	809.3	810.6	811.9
610	813.3	814.6	815.9	817.3	818.6	819.9	821.3	822.6	823.9	825.3
620	826.6	827.9	829.3	830.6	831.9	833.3	834.6	835.9	837.3	838.6
630	839.9	841.3	842.6	843.9	845.3	846.6	847.9	849.3	850.6	851.9
640	853.3	854.6	855.9	857.3	858.6	859.9	861.3	862.6	863.9	865.3
650	866.6	867.9	869.3	870.6	871.9	873.3	874.6	875.9	877.3	878.6
660	879.9	881.3	882.6	883.9	885.3	886.6	887.9	889.3	890.6	891.9
670	893.3	894.6	895.9	897.3	898.6	899.9	901.3	902.6	903.9	905.3
680	906.6	907.9	909.3	910.6	911.9	913.3	914.6	915.9	917.3	918.6
690	919.9	921.3	922.6	923.9	925.3	926.6	927.9	929.3	930.6	931.9
700	933.3	934.6	935.9	937.3	938.6	939.9	941.3	942.6	943.9	945.3
710	946.6	947.9	949.3	950.6	951.9	953.3	954.6	955.9	957.3	958.6
720	959.9	961.3	962.6	963.9	965.3	966.6	967.9	969.3	970.6	971.9
730	973.3	974.6	975.9	977.3	978.6	979.9	981.3	982.6	983.9	985.3
740	986.6	987.9	989.3	990.6	991.9	993.3	994.6	995.9	997.3	998.6
750	999.9	1001.3	1002.6	1003.9	1005.3	1006.6	1007.9	1009.3	1010.6	1011.9
760	1013.3	1014.6	1015.9	1017.2	1018.6	1019.9	1021.2	1022.6	1023.9	1025.2
770	1026.6	1027.9	1029.2	1030.6	1031.9	1033.2	1034.6	1035.9	1037.2	1038.6
780	1039.9	1041.2	1042.6	1043.9	1045.2	1046.6	1047.9	1049.2	1050.6	1051.9
790	1053.2	1054.6	1055.9	1057.2	1058.6	1059.9	1061.2	1062.6	1063.9	1065.2

Table 12

Pressure of Aqueous Vapor over Ice

Temp.	.0	.1	.2	.3	.4	.5	.6	.7	.8	.9
C.	mb.	mb.	mb.	mb.	mb.	mb.	mb.	mb.	mb.	mb.
—35°	0.225	0.222	0.220	0.218	0.215	0.213	0.211	0.208	0.206	0.204
—34	0.251	0.248	0.245	0.243	0.240	0.237	0.235	0.232	0.230	0.227
—33	0.279	0.276	0.273	0.270	0.267	0.265	0.262	0.259	0.256	0.253
—32	0.311	0.307	0.304	0.301	0.298	0.295	0.291	0.288	0.285	0.282
—31	0.345	0.342	0.338	0.335	0.331	0.328	0.324	0.321	0.317	0.314
—30	0.384	0.380	0.376	0.372	0.368	0.364	0.360	0.357	0.353	0.349
—29	0.426	0.421	0.417	0.413	0.408	0.404	0.400	0.396	0.392	0.388
—28	0.472	0.467	0.462	0.458	0.453	0.448	0.444	0.439	0.435	0.430
—27	0.523	0.518	0.512	0.507	0.502	0.497	0.492	0.487	0.482	0.477
—26	0.579	0.573	0.567	0.561	0.556	0.550	0.545	0.539	0.534	0.528
—25	0.640	0.634	0.627	0.621	0.615	0.609	0.602	0.596	0.590	0.585
—24	0.707	0.700	0.693	0.686	0.679	0.673	0.666	0.659	0.653	0.646
—23	0.780	0.773	0.765	0.758	0.750	0.743	0.736	0.728	0.721	0.714
—22	0.861	0.852	0.844	0.836	0.828	0.820	0.812	0.804	0.796	0.788
—21	0.949	0.939	0.930	0.921	0.912	0.904	0.895	0.886	0.878	0.869
—20	1.04	1.03	1.02	1.01	1.00	1.00	0.986	0.976	0.967	0.958
—19	1.15	1.14	1.13	1.12	1.11	1.10	1.09	1.07	1.06	1.05
—18	1.26	1.25	1.24	1.23	1.22	1.20	1.19	1.18	1.17	1.16
—17	1.39	1.37	1.36	1.35	1.34	1.32	1.31	1.30	1.29	1.27
—16	1.52	1.51	1.49	1.48	1.47	1.45	1.44	1.43	1.41	1.40
—15	1.67	1.65	1.64	1.62	1.61	1.59	1.58	1.57	1.55	1.54
—14	1.83	1.81	1.80	1.78	1.76	1.75	1.73	1.72	1.70	1.69
—13	2.00	1.99	1.97	1.95	1.93	1.92	1.90	1.88	1.86	1.85
—12	2.19	2.17	2.15	2.13	2.12	2.10	2.08	2.06	2.04	2.02
—11	2.40	2.38	2.35	2.33	2.31	2.29	2.27	2.25	2.23	2.21
—10	2.62	2.60	2.57	2.55	2.53	2.51	2.48	2.46	2.44	2.42
— 9	2.86	2.83	2.81	2.78	2.76	2.74	2.71	2.69	2.67	2.64
— 8	3.12	3.09	3.07	3.04	3.01	2.99	2.96	2.93	2.91	2.88
— 7	3.40	3.37	3.34	3.31	3.29	3.26	3.23	3.20	3.17	3.15
— 6	3.70	3.67	3.64	3.61	3.58	3.55	3.52	3.49	3.46	3.43
— 5	4.03	4.00	3.97	3.93	3.90	3.87	3.83	3.80	3.77	3.74
— 4	4.39	4.35	4.31	4.28	4.24	4.21	4.17	4.14	4.10	4.07
— 3	4.77	4.73	4.69	4.65	4.61	4.58	4.54	4.50	4.46	4.42
— 2	5.18	5.14	5.10	5.06	5.01	4.97	4.93	4.89	4.85	4.81
— 1	5.63	5.58	5.53	5.49	5.44	5.40	5.36	5.31	5.27	5.23
— 0	6.11	6.06	6.01	5.96	5.91	5.86	5.81	5.77	5.72	5.67

Table 12—*Continued*

Pressure of Aqueous Vapor over Water

Temp.	.0	.1	.2	.3	.4	.5	.6	.7	.8	.9
C.	mb.	mb.	mb.	mb.	mb.	mb.	mb.	mb.	mb.	mb.
0°	6.11	6.15	6.20	6.24	6.29	6.33	6.38	6.42	6.47	6.52
1	6.56	6.61	6.66	6.71	6.76	6.81	6.86	6.90	6.95	7.00
2	7.05	7.10	7.16	7.21	7.26	7.31	7.36	7.42	7.47	7.52
3	7.58	7.63	7.68	7.74	7.79	7.85	7.90	7.96	8.02	8.07
4	8.13	8.19	8.25	8.30	8.36	8.42	8.48	8.54	8.60	8.66
5	8.72	8.78	8.84	8.91	8.97	9.03	9.09	9.16	9.22	9.28
6	9.35	9.41	9.48	9.54	9.61	9.68	9.74	9.81	9.88	9.95
7	10.02	10.09	10.16	10.22	10.30	10.37	10.44	10.51	10.58	10.65
8	10.73	10.80	10.87	10.95	11.02	11.10	11.17	11.25	11.32	11.40
9	11.48	11.56	11.64	11.71	11.79	11.87	11.95	12.03	12.12	12.20
10	12.28	12.36	12.44	12.53	12.61	12.70	12.78	12.87	12.95	13.04
11	13.13	13.21	13.30	13.39	13.48	13.57	13.66	13.75	13.84	13.93
12	14.03	14.12	14.21	14.31	14.40	14.50	14.59	14.69	14.78	14.88
13	14.98	15.08	15.18	15.28	15.38	15.48	15.58	15.68	15.78	15.89
14	15.99	16.09	16.20	16.30	16.41	16.51	16.62	16.73	16.84	16.95
15	17.06	17.17	17.28	17.39	17.50	17.61	17.73	17.84	17.96	18.07
16	18.19	18.30	18.42	18.54	18.66	18.78	18.90	19.02	19.14	19.26
17	19.38	19.51	19.63	19.76	19.88	20.01	20.13	20.26	20.39	20.52
18	20.65	20.78	20.91	21.04	21.17	21.31	21.44	21.58	21.71	21.85
19	21.98	22.12	22.26	22.40	22.54	22.68	22.82	22.96	23.11	23.25
20	23.40	23.54	23.69	23.83	23.98	24.13	24.28	24.43	24.58	24.73
21	24.88	25.04	25.19	25.35	25.50	25.66	25.82	25.98	26.14	26.30
22	26.46	26.62	26.78	26.94	27.11	27.27	27.44	27.61	27.78	27.94
23	28.11	28.28	28.46	28.63	28.80	28.98	29.15	29.33	29.51	29.68
24	29.86	30.04	30.22	30.40	30.59	30.77	30.96	31.14	31.33	31.51
25	31.70	31.89	32.08	32.28	32.47	32.66	32.86	33.05	33.25	33.45
26	33.64	33.84	34.04	34.25	34.45	34.65	34.86	35.06	35.27	35.48
27	35.69	35.90	36.11	36.32	36.53	36.75	36.96	37.18	37.40	37.62
28	37.84	38.06	38.28	38.50	38.73	38.95	39.18	39.41	39.64	39.87
29	40.10	40.33	40.56	40.80	41.04	41.27	41.51	41.75	41.99	42.23
30	42.48	42.72	42.97	43.21	43.46	43.71	43.96	44.21	44.47	44.72
31	44.98	45.23	45.49	45.75	46.01	46.27	46.54	46.80	47.07	47.33
32	47.60	47.87	48.14	48.42	48.69	48.97	49.24	49.52	49.80	50.08
33	50.36	50.65	50.93	51.22	51.50	51.79	52.08	52.37	52.67	52.96
34	53.26	53.56	53.85	54.15	54.46	54.76	55.06	55.37	55.68	55.99
35	56.30	56.61	56.92	57.24	57.56	57.87	58.19	58.51	58.84	59.16
36	59.49	59.81	60.14	60.47	60.81	61.14	61.47	61.81	62.15	62.49
37	62.83	63.17	63.52	63.86	64.21	64.56	64.91	65.27	65.62	65.98
38	66.34	66.69	67.06	67.42	67.78	68.15	68.52	68.89	69.26	69.63
39	70.01	70.38	70.76	71.14	71.53	71.91	72.30	72.68	73.07	73.46
40	73.86	74.25	74.65	75.04	75.44	75.85	76.25	76.66	77.06	77.47
41	77.88	78.30	78.71	79.13	79.55	79.97	80.39	80.81	81.24	81.67
42	82.10	82.53	82.97	83.40	83.84	84.28	84.72	85.17	85.61	86.06
43	86.51	86.96	87.42	87.87	88.33	88.79	89.26	89.72	90.19	90.66
44	91.13	91.60	92.07	92.55	93.03	93.51	93.99	94.48	94.97	95.46

Table 13

Calculation of the Potential Temperature (θ)

Multiply the *absolute temperature* by the factor in the table below which corresponds to the existing pressure, to obtain the *Potential Temperature* (θ)

p in Milli-bars	0	10	20	30	40	50	60	70	80	90
0	......	3.767	3.085	2.745	2.528	2.370	2.252	2.151	2.070	2.001
100	1.943	1.888	1.842	1.800	1.762	1.726	1.695	1.666	1.639	1.613
200	1.590	1.568	1.547	1.527	1.501	1.491	1.474	1.458	1.445	1.428
300	1.414	1.401	1.388	1.376	1.365	1.353	1.342	1.332	1.321	1.312
400	1.302	1.293	1.285	1.276	1.267	1.259	1.251	1.244	1.236	1.228
500	1.221	1.214	1.207	1.200	1.194	1.188	1.182	1.176	1.170	1.164
600	1.159	1.153	1.148	1.142	1.137	1.132	1.127	1.122	1.118	1.113
700	1.108	1.104	1.099	1.095	1.091	1.086	1.082	1.078	1.074	1.070
800	1.066	1.062	1.059	1.055	1.051	1.048	1.044	1.041	1.037	1.034
900	1.031	1.028	1.024	1.021	1.018	1.015	1.012	1.009	1.006	1.003
1000	1.000	0.997	0.994	0.991	0.988	0.985	0.982	0.979	0.976	0.973

Table for $\left(\frac{p_o}{p}\right)^{.288}$ (where p_o = 1000 mb.)

Table 14

Calculation of the Partial Potential Temperature (θ_d)

(Add the quantities below to the Potential Temperature (θ) to obtain the Partial Temperature of the Dry Air (θ_d).)

W \ θ	250	260	270	280	290	300	310	320	330	340	350
0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
1	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.2	0.2	0.2
2	0.2	0.2	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3
3	......	0.4	0.4	0.4	0.4	0.4	0.4	0.4	0.5	0.5	0.5
4	......	0.5	0.5	0.5	0.5	0.6	0.6	0.6	0.6	0.6	0.7
5	......	0.6	0.6	0.7	0.7	0.7	0.7	0.7	0.8	0.8	0.8
6	......	......	0.8	0.8	0.8	0.8	0.9	0.9	0.9	1.0	1.0
7	......	......	0.9	0.9	0.9	1.0	1.0	1.0	1.1	1.1	1.1
8	......	......	1.0	1.0	1.1	1.1	1.2	1.2	1.2	1.3	1.3
9	......	......	1.1	1.2	1.2	1.3	1.3	1.3	1.4	1.4	1.5
10	......	......	1.3	1.3	1.3	1.4	1.4	1.5	1.5	1.5	1.6
11	......	......	1.4	1.4	1.5	1.5	1.6	1.6	1.7	1.7	1.8
12	......	......	.....	1.6	1.6	1.7	1.7	1.8	1.8	1.9	1.9
13	......	......	.....	1.7	1.7	1.8	1.9	1.9	2.0	2.0	2.1
14	......	......	.....	1.8	1.9	1.9	2.0	2.1	2.1	2.2	2.3
15	......	......	.....	1.9	2.0	2.1	2.2	2.2	2.3	2.4	2.4
16	......	......	.....	2.1	2.1	2.2	2.3	2.4	2.4	2.5	2.6
17	......	......	.....	2.2	2.3	2.4	2.4	2.5	2.6	2.7	2.7
18	......	......	.....	2.3	2.4	2.5	2.6	2.7	2.7	2.8	2.9
19	......	......	.....	2.5	2.5	2.6	2.7	2.8	2.9	3.0	3.1
20	......	......	.....	2.6	2.7	2.8	2.9	2.9	3.0	3.1	3.2

TABLE 15

CALCULATION OF THE EQUIVALENT POTENTIAL TEMPERATURE (θ_e)

Add the quantities below to the Partial Potential Temperature (θ_d) to obtain the Equivalent Potential Temperature (θ_e)

W \ θ_d	250	260	270	280	290	300	310	320	330	340	350
0.0	0	0	0	0	0	0	0	0	0	0	0
1.0	2	3	3	3	3	3	3	3	4	4	4
2.0	......	5	5	6	6	6	6	7	7	7	7
3.0	......	......	8	8	8	9	9	9	10	10	11
4.0	......	......	10	10	11	12	12	12	13	13	14
5.0	......	......	12	13	14	14	15	15	16	17	17
6.0	......	......	.....	15	16	17	18	18	19	20	21
7.0	......	......	.....	18	19	19	20	21	22	23	24
8.0	......	......	.....	20	21	22	23	24	25	26	27
9.0	......	......	.....	23	24	25	26	27	28	29	30
10.0	......	......	.....	25	26	27	29	30	31	32	33
11.0	......	......	.....	.....	29	30	31	33	34	35	36
12.0	......	......	.....	.....	31	33	34	35	37	38	39
13.0	......	......	.....	.....	34	35	37	38	40	42	43
14.0	......	......	.....	.....	37	38	39	41	43	45	47
15.0	......	......	.....	.....	39	40	42	44	46	48	50
16.0	......	......	.....	.....	41	43	45	47	49	51	53
17.0	......	......	.....	.....	44	46	48	50	52	54	56
18.0	......	......	.....	.....	46	48	50	53	55	57	60
19.0	......	......	.....	.....	49	51	53	56	58	60	63
20.0	......	......	.....	.....	51	53	56	58	61	63	66

TABLE 16

THE STANDARD ATMOSPHERE

	Metric	English
Standard temperature..........	15° C.	59° F.
Standard temperature absolute..	288° A.	
Standard pressure..............	760 mm. of Hg.	29.92117 in. of Hg.
Standard pressure..............	10332.276 kg/m²	2116.229 lb/ft²
Standard gravity..............	9.80665 m/sec²	32.1740 ft/sec²
Standard specific weight........	1.2255 kg/m³	0.07651 lb/ft³
Standard density..............	0.124966 kg/m/sec	0.002378 lb/ft/sec
Standard temperature gradient..	0.0065° C./m	0.00356617° F./ft
Standard isothermal temperature	−55° C.	−67° F.
Standard gas constant for air....	29.2708	53.33089

TABLE 17

DISTANCE IN DYNAMIC METERS BETWEEN PRINCIPAL ISOBARIC SURFACES FOR VARIOUS VALUES OF THE VIRTUAL TEMPERATURE (T_v) *

	Mean T_v (C°)	0	1	2	3	4	5	6	7	8	9
100 mb											
	−100	3442	3422	3402	3382	3362	3343	3323	3303	3283	3263
	− 90	3641	3621	3601	3581	3561	3542	3522	3502	3482	3462
	− 80	3840	3820	3800	3780	3760	3741	3721	3701	3681	3661
	− 70	4039	4019	3999	3979	3959	3939	3920	3900	3880	3860
	− 60	4238	4218	4198	4178	4158	4138	4119	4099	4079	4059
	− 50	4437	4417	4397	4377	4357	4337	4318	4298	4278	4258
	− 40	4636	4616	4596	4576	4556	4536	4516	4497	4477	4457
	− 30	4835	4815	4795	4775	4755	4735	4715	4696	4676	4656
200 mb											
	− 90	2130	2118	2107	2095	2083	2072	2060	2048	2037	2025
	− 80	2246	2235	2223	2211	2200	2188	2176	2165	2153	2142
	− 70	2363	2351	2339	2328	2316	2304	2295	2281	2270	2258
	− 60	2479	2467	2456	2444	2432	2421	2409	2398	2386	2374
	− 50	2595	2584	2572	2561	2549	2537	2526	2514	2502	2491
	− 40	2712	2700	2689	2677	2665	2654	2642	2630	2619	2607
	− 30	2828	2817	2805	2793	2782	2770	2758	2747	2735	2723
	− 20	2945	2933	2921	2910	2898	2886	2875	2863	2851	2840
300 mb											
	− 80	1594	1585	1577	1569	1561	1552	1544	1536	1528	1519
	− 70	1676	1668	1660	1652	1643	1635	1627	1619	1610	1602
	− 60	1759	1751	1742	1743	1726	1718	1709	1701	1693	1685
	− 50	1841	1833	1825	1817	1808	1800	1792	1784	1775	1767
	− 40	1924	1916	1908	1899	1891	1883	1874	1866	1858	1850
	− 30	2007	1998	1990	1982	1974	1965	1957	1949	1941	1932
	− 20	2089	2081	2073	2064	2056	2048	2040	2031	2023	2015
	− 10	2172	2164	2155	2147	2139	2130	2122	2114	2106	2097
400 mb											
	− 70	1300	1294	1287	1281	1275	1268	1262	1255	1249	1243
	− 60	1364	1358	1351	1345	1339	1332	1326	1319	1313	1307
	− 50	1428	1422	1416	1409	1403	1396	1390	1384	1377	1371
	− 40	1492	1486	1480	1473	1467	1460	1454	1448	1441	1435
	− 30	1556	1550	1544	1537	1531	1524	1518	1512	1505	1499
	− 20	1621	1614	1608	1601	1595	1588	1582	1576	1569	1563
	− 10	1685	1678	1672	1665	1659	1653	1646	1640	1633	1627
	− 0	1749	1742	1736	1729	1723	1717	1710	1704	1697	1691
500 mb											
	− 60	1115	1109	1104	1099	1094	1089	1083	1078	1073	1068
	− 50	1167	1162	1157	1151	1146	1141	1136	1130	1125	1120
	− 40	1219	1214	1209	1204	1198	1193	1188	1183	1178	1172
	− 30	1272	1266	1261	1256	1251	1246	1240	1235	1230	1225
600 mb											
		0	1	2	3	4	5	6	7	8	9

* $T_v = \dfrac{T}{1 - \frac{3}{8}\cdot\frac{e}{p}}$ where T_v is the virtual temperature, T the absolute temperature, e the vapor pressure, and p the total pressure.

TABLE 17—*Continued*

DISTANCE IN DYNAMIC METERS BETWEEN PRINCIPAL ISOBARIC SURFACES FOR VARIOUS VALUES OF THE VIRTUAL TEMPERATURE (T_v)

	Mean T_v (C°)	0	1	2	3	4	5	6	7	8	9
500 mb											
	− 20	1324	1319	1314	1308	1303	1298	1293	1287	1282	1277
	− 10	1376	1371	1366	1361	1355	1350	1345	1340	1335	1329
	− 0	1429	1423	1418	1413	1408	1403	1397	1392	1387	1382
	+ 0	1429	1434	1439	1444	1450	1455	1460	1465	1471	1476
600 mb											
	− 50	987	982	978	973	969	965	960	956	951	947
	− 40	1031	1027	1022	1018	1013	1009	1004	1000	996	991
	− 30	1075	1071	1066	1062	1058	1053	1049	1044	1040	1035
	− 20	1119	1115	1111	1106	1102	1097	1093	1088	1084	1080
	− 10	1164	1159	1155	1150	1146	1142	1137	1133	1128	1124
	− 0	1208	1204	1199	1195	1190	1186	1181	1177	1173	2268
	+ 0	1208	1212	1217	1221	1226	1230	1235	1239	1243	1248
	10	1252	1257	1261	1265	1270	1274	1279	1283	1288	1292
700 mb											
	− 40	893	889	885	882	878	874	870	866	862	859
	− 30	931	928	924	920	916	912	908	905	901	897
	− 20	970	966	962	958	954	951	947	943	939	935
	− 10	1008	1004	1000	997	993	989	985	981	977	974
	− 0	1046	1043	1039	1035	1031	1027	1023	1020	1016	1012
	+ 0	1046	1050	1054	1058	1062	1066	1069	1073	1077	1081
	10	1085	1089	1092	1096	1100	1104	1108	1112	1115	1119
	20	1123	1127	1131	1135	1138	1142	1146	1150	1154	1158
800 mb											
	− 40	788	784	781	778	774	771	767	764	761	757
	− 30	822	818	815	811	808	805	801	798	795	791
	− 20	855	852	849	845	842	838	835	832	828	825
	− 10	889	886	882	879	876	872	869	866	862	859
	− 0	923	920	916	913	909	906	903	899	896	893
	+ 0	923	926	930	933	937	940	943	947	950	953
	10	957	960	964	967	970	974	977	980	984	987
	20	991	994	997	1001	1004	1008	1011	1014	1018	1021
	30	1024	1028	1031	1035	1038	1041	1045	1048	1051	1055
900 mb											
	− 40	705	702	699	696	693	690	687	684	680	677
	− 30	735	732	729	726	723	720	714	714	711	708
	− 20	765	762	759	756	753	750	747	744	741	738
	− 10	795	792	789	786	783	780	777	774	771	768
	− 0	826	823	820	817	814	811	808	804	801	798
	+ 0	826	829	832	835	838	841	844	847	850	853
	10	856	859	862	865	868	871	874	877	880	883
	20	886	889	892	895	898	901	904	907	910	913
	30	916	919	922	925	928	931	935	938	941	944
	40	947	950	953	956	959	962	965	968	971	974
1000 mb											
		0	1	2	3	4	5	6	7	8	9

TABLE 18

NUMBER OF SOLENOIDS INCLUDED WITHIN A CLOSED CURVE, BOUNDED BY THE 1000 MB. ISOBAR (P_0), THE ISOBAR P_1, AND THE TEMPERATURES T_a AND T_b (° C.) (AFTER BJERKNES)

P_1 \ Ta-Tb	1°	5°	10°	20°	30°	40°
100	660.8	3304	6608	13217	19825	26434
200	461.9	2310	4619	9238	13857	18476
300	345.5	1728	3455.	6910	10365	13820
400	262.9	1315	2629	5258	7887	10518
500	198.9	995	1989	3978	4967	7958
600	146.9	735	1469	2938	4407	5876
700	102.5	513	1025	2050	3075	4100
800	63.0	315	630	1260	1890	2520
900	30.1	151	301	602	903	1204
1000	0	0	0	0	0	0

TABLE 19

LATENT HEAT OF VAPORIZATION OF WATER (L), IN GRAM-CALORIES PER GRAM

° C. L	0 594.9	5 592.4	10 590.0	15 587.5	20 585.0	25 582.4	30 579.8
° C. L	40 574.5	50 569.0	60 563.4	70 557.6	80 551.6	90 545.5	100 539.1

Table 20

Temperature of the Condensation Level (° C.)

(This table may be used to calculate the lift necessary to saturate a sample of air. The difference between the air temperature and the temperature of the condensation level gives the number of hundred meters lift required to reach saturation.)

θ_e \ W	0	0.1	0.2	0.3	0.4	0.5	0.6	0.7	0.8	0.9	1.0
240	....	−42.8	34.7	29.5	26.0	23.4	21.0	19.0	17.0	15.3	14.0
242	....	−43.1	35.0	29.8	26.4	23.8	21.4	19.4	17.5	15.7	14.3
244	....	−43.4	35.3	30.4	26.8	24.2	21.8	19.8	17.9	16.1	14.7
246	....	−43.7	35.7	30.8	27.2	24.5	22.2	20.2	18.3	16.5	15.0
248	....	−44.0	36.1	31.2	27.6	24.8	22.6	20.6	18.7	16.9	15.4
250	....	−44.3	36.5	31.6	28.0	25.7	23.0	21.0	19.0	17.3	15.8
252	....	−44.6	36.9	32.0	28.4	25.6	23.4	21.3	19.4	17.7	16.2
254	....	−44.9	37.0	32.4	28.7	25.9	23.7	21.7	19.8	18.1	16.6
256	....	−45.2	37.4	32.7	29.0	26.3	24.0	22.0	20.2	18.5	17.0
258	....	−45.5	37.7	33.0	29.4	26.6	24.3	22.3	20.5	18.8	17.4
260	....	−45.8	38.0	33.4	29.8	26.9	24.7	22.6	20.9	19.1	17.8
262	....	−46.1	38.3	33.7	30.1	27.2	25.0	23.0	21.2	19.5	18.2
264	....	−46.4	38.6	34.0	30.4	27.6	25.4	23.4	21.6	19.9	18.5
266	....	−46.8	38.9	34.3	30.7	27.9	25.7	23.7	22.0	20.3	18.8
268	....	−47.1	39.2	34.6	31.0	28.2	26.0	24.0	22.3	20.7	19.2
270	....	−47.4	39.5	34.9	31.3	28.5	26.3	24.4	22.6	21.0	19.6
272	....	−47.7	39.8	35.1	31.6	28.8	26.6	24.7	23.0	21.4	20.0
274	....	−48.0	40.1	35.3	31.9	29.1	26.9	25.0	23.3	21.8	20.3
276	....	−48.3	40.4	35.6	32.2	29.4	27.2	25.3	23.7	22.2	20.7
278	....	−48.6	40.7	35.9	32.5	29.7	27.5	25.6	24.0	22.5	21.0
280	....	−49.0	41.0	36.1	32.8	30.0	27.8	25.9	24.3	22.8	21.3
282	....	−49.3	41.2	36.4	33.1	30.3	28.1	26.2	24.6	23.1	21.7
284	....	−49.6	41.5	36.7	33.4	30.6	28.4	26.5	24.9	23.4	22.1
286	....	−49.9	41.7	37.0	33.7	30.9	28.7	26.8	25.2	23.7	22.4
288	....	−50.2	42.0	37.2	34.0	31.2	29.0	27.1	25.5	24.0	22.7
290	....	−50.5	42.2	37.5	34.3	31.5	29.3	27.4	25.8	24.2	23.0
292	....	−50.8	42.5	37.8	34.6	31.8	29.6	27.7	26.1	24.5	23.3
294	....	−51.2	42.7	38.1	34.9	32.1	29.9	28.0	26.4	24.8	23.6
296	....	−51.5	43.0	38.4	35.2	32.4	30.2	28.3	26.7	25.0	23.9
298	....	−51.8	43.2	38.7	35.5	32.7	30.4	28.6	27.0	25.3	24.2
300	....	−52.1	43.5	39.0	35.7	33.0	30.7	28.9	27.3	25.6	24.5
302	....	−52.4	43.7	39.3	36.0	33.3	31.0	29.1	27.6	25.9	24.8
304	....	−52.7	44.0	39.6	36.2	33.6	31.3	29.4	27.9	26.2	25.0
306	....	−53.0	44.2	39.9	36.5	33.9	31.6	29.7	28.1	26.5	25.2
308	....	−53.3	44.5	40.2	36.7	34.2	31.9	30.0	28.3	26.8	25.5
310	....	−53.6	44.7	40.4	37.0	34.5	32.2	30.2	28.6	27.1	25.8
312	....	−53.9	45.0	40.7	37.2	34.8	32.4	30.5	28.8	27.4	26.1
314	....	−54.2	45.2	40.9	37.5	35.0	32.7	30.8	29.1	27.7	26.4
316	....	−54.5	45.5	41.2	37.7	35.2	33.0	31.0	29.3	28.0	26.7
318	....	−54.9	45.7	41.5	37.9	35.5	33.3	31.3	29.5	28.2	27.0
320	....	−55.2	46.0	41.8	38.1	35.8	33.6	31.5	29.8	28.5	27.2
322	....	−55.5	46.3	42.0	38.4	36.0	33.8	31.8	30.0	28.8	27.5
324	....	−55.8	46.6	42.2	38.7	36.3	34.0	32.1	30.3	29.0	27.8
326	....	−56.1	46.9	42.4	39.0	36.5	34.3	32.4	30.6	29.2	28.1
328	....	−56.4	47.1	42.6	39.2	36.7	34.5	32.7	30.8	29.4	28.4
330	....	−56.7	47.4	42.8	39.5	36.9	34.8	33.0	31.1	29.7	28.6
332	....	−57.0	47.6	43.0	39.8	37.1	35.0	33.3	31.3	30.0	28.8
334	....	−57.4	47.9	43.2	40.0	37.4	35.2	33.5	31.5	30.2	29.0
336	....	−57.7	48.1	43.4	40.3	37.6	35.5	33.8	31.8	30.5	29.3
338	....	−58.0	48.4	43.6	40.5	37.8	35.7	34.0	32.1	30.7	29.6
340	....	−58.3	48.7	43.8	40.7	38.0	36.0	34.3	32.4	31.0	29.9
342	....	−58.6	49.0	44.0	41.0	38.3	36.2	34.5	32.7	31.2	30.2
344	....	−58.9	49.2	44.2	41.2	38.5	36.4	34.8	33.0	31.5	30.4
346	....	−59.2	49.5	44.4	41.5	38.8	36.6	35.0	33.2	31.7	30.7
348	....	−59.5	49.7	44.7	41.7	39.0	36.8	35.3	33.5	32.0	30.9
θ_e / W	0	0.1	0.2	0.3	0.4	0.5	0.6	0.7	0.8	0.9	1.0

TABLE 20—*Continued*

TEMPERATURE OF THE CONDENSATION LEVEL (° C.)—*Continued*

θ_e \ W	1.0	1.5	2.0	2.5	3.0	3.5	4.0	4.5	5.0	5.5	6.0	6.5	7.0
250	−15.8	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
252	−16.2	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
254	−16.6	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
256	−17.0	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
258	−17.4	11.7	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
260	−17.8	12.0	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
262	−18.2	12.4	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
264	−18.5	12.7	− 7.5	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
266	−18.8	13.0	− 7.9	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
268	−19.2	13.4	− 8.3	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
270	−19.6	13.7	− 8.7	− 4.7	.....	.....	.....	.....	.....	.....	.....	.....	.....
272	−20.0	14.1	− 9.0	− 5.1	− 2.1	.....	.....	.....	.....	.....	.....	.....	.....
274	−20.3	14.5	− 9.3	− 5.5	− 2.5	.....	.....	.....	.....	.....	.....	.....	.....
276	−20.7	14.9	− 9.6	− 5.9	− 2.9	− 0.1	.....	.....	.....	.....	.....	.....	.....
278	−21.0	15.3	10.0	− 6.3	− 3.3	− 0.6	+ 1.9	.....	.....	.....	.....	.....	.....
280	−21.3	15.6	10.4	− 6.7	− 3.7	− 1.0	+ 1.5	.....	.....	.....	.....	.....	.....
282	−21.7	16.0	10.8	− 7.1	− 4.1	− 1.4	+ 1.2	+ 3.4	.....	.....	.....	.....	.....
284	−22.1	16.4	11.1	− 7.5	− 4.4	− 1.8	+ 0.8	+ 3.0	+5.0	.....	.....	.....	.....
286	−22.4	16.7	11.5	− 7.9	− 4.8	− 2.1	+ 0.4	+ 2.6	+4.6	.....	.....	.....	.....
288	−22.7	17.0	11.9	− 8.3	− 5.2	− 2.5	0.0	+ 2.2	+4.2	+6.0	.....	.....	.....
290	−23.0	17.4	12.3	− 8.7	− 5.6	− 2.9	− 0.4	+ 1.8	+3.8	+5.7	+7.4	.....	.....
292	−23.3	17.7	12.7	− 9.0	− 6.0	− 3.3	− 0.8	+ 1.4	+3.4	+5.3	+7.0	+8.6	.....
294	−23.6	18.1	13.0	− 9.4	− 6.4	− 3.7	− 1.2	+ 1.0	+3.0	+4.9	+6.6	+8.2	+9.8
296	−23.9	18.4	13.4	− 9.7	− 6.8	− 4.0	− 1.6	+ 0.6	+2.6	+4.5	+6.2	+7.9	+9.4
298	−24.2	18.7	13.7	10.0	− 7.1	− 4.3	− 2.0	+ 0.2	+2.3	+4.1	+5.8	+7.5	+9.0
300	−24.5	19.0	14.0	10.4	− 7.5	− 4.7	− 2.4	− 0.2	+1.9	+3.7	+5.4	+7.1	+8.6
302	−24.8	19.3	14.4	10.8	− 7.9	− 5.1	− 2.7	− 0.6	+1.6	+3.3	+5.0	+6.7	+8.2
304	−25.0	19.6	14.7	11.1	− 8.2	− 5.5	− 3.1	− 1.0	+1.2	+2.9	+4.6	+6.3	+7.8
306	−25.2	20.0	15.0	11.5	− 8.5	− 5.9	− 3.5	− 1.4	+0.8	+2.5	+4.2	+5.9	+7.4
308	−25.5	20.3	15.4	11.9	− 8.8	− 6.3	− 3.9	− 1.8	+0.4	+2.1	+3.8	+5.5	+7.0
310	−25.8	20.6	15.7	12.2	− 9.1	− 6.7	− 4.3	− 2.1	0.0	+1.8	+3.5	+5.1	+6.6
312	−26.1	20.9	16.0	12.5	− 9.5	− 7.0	− 4.6	− 2.5	−0.4	+1.4	+3.1	+4.7	+6.3
314	−26.4	21.2	16.4	12.8	− 9.9	− 7.3	− 4.9	− 2.8	−0.8	+1.0	+2.8	+4.3	+5.9
316	−26.7	21.5	16.7	13.1	10.3	− 7.7	− 5.2	− 3.1	−1.1	+0.7	+2.4	+4.0	+5.5
318	−27.0	21.8	17.0	13.4	10.6	− 8.0	− 5.6	− 3.5	−1.5	+0.3	+2.0	+3.6	+5.1
320	−27.2	22.1	17.2	13.7	10.9	− 8.3	− 6.0	− 3.9	−1.8	0.0	+1.7	+3.2	+4.8
322	−27.5	22.4	17.5	14.0	11.1	− 8.6	− 6.3	− 4.2	−2.2	−0.3	+1.3	+2.9	+4.5
324	−27.8	22.7	17.8	14.2	11.4	− 9.0	− 6.7	− 4.5	−2.5	−0.7	+1.0	+2.6	+4.1
326	−28.1	23.0	18.1	14.5	11.8	− 9.3	− 7.0	− 4.8	−2.9	−1.0	+0.6	+2.2	+3.7
328	−28.4	23.3	18.4	14.8	12.1	− 9.6	− 7.3	− 5.1	−3.2	−1.4	+0.2	+1.8	+3.3
330	−28.6	23.6	18.7	15.1	12.4	− 9.9	− 7.6	− 5.5	−3.6	−1.8	−0.1	+1.5	+3.0
332	−28.8	23.9	18.9	15.4	12.7	10.2	− 8.0	− 5.9	−4.0	−2.2	−0.4	+1.1	+2.7
334	−29.0	24.2	19.2	15.7	13.0	10.5	− 8.3	− 6.2	−4.3	−2.5	−0.7	+0.8	+2.4
336	−29.3	24.5	19.5	16.0	13.3	10.8	− 8.6	− 6.5	−4.6	−2.8	−1.0	+0.5	+2.0
338	−29.6	24.7	19.8	16.3	13.6	11.1	− 8.9	− 6.8	−4.9	−3.1	−1.4	+0.1	+1.6
340	−29.9	24.9	20.1	16.6	13.9	11.4	− 9.2	− 7.1	−5.2	−3.5	−1.8	−0.3	+1.3
342	−30.2	25.1	20.4	16.9	14.2	11.7	− 9.5	− 7.5	−5.5	−3.8	−2.1	−0.6	+1.0
344	−30.4	25.4	20.6	17.2	14.5	12.0	− 9.8	− 7.8	−5.9	−4.1	−2.4	−0.9	+0.7
346	−30.7	25.6	20.8	17.5	14.8	12.3	10.1	− 8.1	−6.2	−4.4	−2.7	−1.2	+0.3
348	−30.9	25.9	21.0	17.8	15.0	12.6	10.4	− 8.4	−6.5	−4.7	−3.0	−1.5	0.0
350	−31.1	26.1	21.2	18.1	15.3	12.9	10.7	− 8.7	−6.8	−5.0	−3.3	−1.9	−0.4
352	.....	−26.4	21.5	18.4	15.6	13.2	11.0	− 9.0	−7.1	−5.3	−3.7	−2.2	−0.7
354	.....	.....	−21.7	18.7	15.9	13.5	11.3	− 9.3	−7.4	−5.7	−4.0	−2.5	−1.0
356	.....	.....	.....	−19.0	16.2	13.8	11.6	− 9.6	−7.8	−6.0	−4.3	−2.8	−1.4
358	.....	.....	.....	.....	−16.5	14.0	11.9	− 9.9	−8.1	−6.3	−4.6	−3.1	−1.7
360	.....	.....	.....	.....	.....	−14.5	12.1	10.2	−8.4	−6.6	−4.9	−3.4	−2.0
362	.....	.....	.....	.....	.....	.....	.....	−10.5	−8.7	−6.9	−5.2	−3.7	−2.3
364	.....	.....	.....	.....	.....	.....	.....	.....	.....	−7.2	−5.5	−4.0	−2.6
366	.....	.....	.....	.....	.....	.....	.....	.....	.....	−7.5	−5.8	−4.3	−2.9
368	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	−6.1	−4.6	−3.2
θ_e / W	1.0	1.5	2.0	2.5	3.0	3.5	4.0	4.5	5.0	5.5	6.0	6.5	7.0

TABLE 20—*Continued*

TEMPERATURE OF THE CONDENSATION LEVEL (° C.)—*Continued*

7.5	8.0	8.5	9.0	9.5	10.0	10.5	11.0	11.5	12.0	12.5	13.0	13.5	W / θ_e
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	250
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	252
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	254
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	256
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	258
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	260
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	262
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	264
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	266
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	268
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	270
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	272
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	274
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	276
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	278
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	280
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	282
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	284
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	286
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	288
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	290
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	292
.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	294
+11.0	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	296
+10.6	11 9	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	298
+10.2	11.5	12.9	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	300
+9.7	11.2	12.5	13.8	.....	.....	.....	.....	.....	.....	.....	.....	.....	302
+9.3	10 8	12.1	13.4	.....	.....	.....	.....	.....	.....	.....	.....	.....	304
+8.9	10 3	11.7	13.0	14.3	.....	.....	.....	.....	.....	.....	.....	.....	306
+8.5	+9.9	11.3	12.6	13.9	15.2	.....	.....	.....	.....	.....	.....	.....	308
+8.1	+9.5	10.9	12.2	13.5	14.7	15.8	17.0	.....	.....	.....	.....	.....	310
+7.7	+9.1	10.5	11.8	13.1	14.3	15.4	16.6	.....	.....	.....	.....	.....	312
+7.3	+8.7	10.1	11.4	12.7	13.9	15.0	16.2	17.3	18.3	.....	.....	.....	314
+7.0	+8.3	+9.7	11.0	12.3	13.5	14.6	15.8	16.9	17.9	19.2	.....	.....	316
+6.6	+7.9	+9.3	10.6	11.9	13.1	14.2	15.4	16.5	17.5	18.6	19.6	.....	318
+6.2	+7.5	+9.0	10.2	11.6	12.7	13.8	15.0	16.1	17.2	18.2	19.2	20.0	320
+5.8	+7.2	+8.6	+9.8	11.2	12.4	13.4	14.6	15.7	16.8	17.8	18.8	19.7	322
+5.4	+6.8	+8.2	+9.4	10.8	12.0	13.0	14.2	15.3	16.4	17.4	18.4	19.4	324
+5.1	+6.5	+7.8	+9.0	10.4	11.6	12.7	13.8	14.9	16.0	17.0	18 0	19.0	326
+4.7	+6.1	+7.4	+8.7	10.0	11.2	12.4	13.4	14.5	15.6	16.6	17.6	18.6	328
+4.3	+5.8	+7.0	+8.4	+9.6	10.8	12.0	13.0	14.1	15.2	16.2	17.2	18 2	330
+4.0	+5.4	+6.6	+8.0	+9.2	10.4	11.6	12.6	13.7	14.8	15.8	16.8	17.8	332
+3.7	+5.0	+6.3	+7.8	+8.8	10.0	11.2	12.3	13.4	14.4	15.5	16.5	17.5	334
+3.4	+4.6	+5.9	+7.2	+8.5	+9.6	10.8	11.9	13.0	14.0	15.1	16.1	17.1	336
+3.0	+4.3	+5.6	+6.9	+8.1	+9.2	10.4	11.5	12.6	13.6	14.7	15.7	16.7	338
+2.7	+4.0	+5.3	+6.5	+7.7	+8.9	10.0	11.1	12.2	13.2	14.3	15.3	16.3	340
+2.3	+3.7	+5.0	+6.1	+7.4	+8.6	+9.6	10.7	11.8	12.9	14.0	15.0	15.9	342
+2.0	+3.4	+4.7	+5.7	+7.0	+8.2	+9.3	10.4	11.5	12.6	13.6	14.6	15.5	344
+1.7	+3.0	+4.3	+5.4	+6.7	+7.8	+9.0	10.0	11.1	12.2	13.2	14.2	15.2	346
+1.3	+2.7	+4.0	+5.1	+6.3	+7.5	+8.6	+9.6	10.7	11.8	12.8	13.8	14.8	348
+1.0	+2.3	+3.7	+4.8	+6.0	+7.2	+8.2	+9.3	10 3	11.5	12.4	13.4	14.4	350
+0.7	+2.0	+3.3	+4.5	+5.7	+6.8	+7.8	+9.0	10.0	11.1	12.0	13.0	14 0	352
+0.4	+1.6	+3.0	+4.2	+5.4	+6.5	+7.5	+8.7	+9.7	10.7	11.7	12.7	13.7	354
0.0	+1.3	+2.7	+3.8	+5.0	+6.2	+7.2	+8.3	+9.3	10.4	11.4	12.4	13.4	356
−3.3	+1.0	+2.3	+3.5	+4.7	+5.9	+6.8	+7.9	+9.0	10.0	11.0	32.0	13.0	358
−0.6	+0.8	+2.0	+3.2	+4.4	+5.6	+6.5	+7.6	+8.7	+9.7	10.7	11.6	12.7	360
−0.9	+0.5	+1.6	+2.9	+4.0	+5.2	+6.2	+7.3	+8.3	+9.4	10.4	11.3	14.2	362
−1.2	+0.1	+1.3	+2.5	+3.7	+4.9	+5.9	+7.0	+8.0	+9.0	10.0	11.0	12.0	364
−1.5	−0.3	+1.0	+2.2	+3.4	+4.6	+5.6	+6.7	+7.7	+8.7	+9.7	10.7	11.6	366
−1.8	−0.5	+0.7	+1.9	+3.0	+4.2	+5.3	+6.4	+7.4	+8.3	+9.4	10.4	11.2	368
7.5	8.0	8.5	9.0	9.5	10.10	10.5	11.0	11.5	12.0	12.5	13.0	13.5	θ_e / W

TABLE 20—*Continued*

TEMPERATURE OF THE CONDENSATION LEVEL (° C.)—*Continued*

θ_e \ W	14.0	14.5	15.0	15.5	16.0	16.5	17.0	17.5	18.0	18.5	19.0	19.5	20.0
322	+20.8	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
324	+20.4	21.4	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
326	+20.0	21.0	21.8	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
328	+19.6	20.6	21.4	22.4	.....	.....	.....	.....	.....	.....	.....	.....	.....
330	+19.2	20.2	21.0	22.0	23.0	.....	.....	.....	.....	.....	.....	.....	.....
332	+18.8	19.8	20.7	21.6	22.5	23.4	.....	.....	.....	.....	.....	.....	.....
334	+18.4	19.3	20.3	21.2	22.1	23.0	23.6	.....	.....	.....	.....	.....	.....
336	+18.0	18.9	19.9	20.8	21.7	22.6	23.3	24.2	25.1	.....	.....	.....	.....
338	+17.6	18.5	19.5	20.4	21.3	22.2	23.0	23.8	24.7	.....	.....	.....	.....
340	+17.3	18.1	19.1	20.0	20.9	21.8	22.6	23.5	24.3	25.1	25.8	.....	....
342	+16.9	17.7	18.7	19.6	20.5	21.4	22.2	23.1	23.9	24.7	25.5	26.3	.....
344	+16.5	17.3	18.3	19.2	20.1	21.0	21.8	22.7	23.6	24.3	25.1	25.9	26.6
346	+16.1	17.0	17.9	18.8	19.7	20.6	21.4	22.3	23.2	23.9	24.7	25.5	26.2
348	+15.7	16.6	17.6	18.5	19.3	20.2	21.0	21.9	22.8	23.5	24.3	25.1	25.8
350	+15.4	16.3	17.2	18.1	19.0	19.8	20.7	21.6	22.4	23.2	23.9	24.8	25.4
352	+15.0	15.9	16.8	17.7	18.6	19.5	20.3	21.2	22.0	22.8	23.6	24.4	25.0
354	+14.6	15.5	16.5	17.3	18.3	19.1	19.9	20.8	21.6	22.4	23.2	24.0	24.7
356	+14.3	15.2	16.1	17.0	17.9	18.7	19.5	20.5	21.2	22.0	22.8	23.6	24.3
358	+13.9	14.8	15.7	16.6	17.5	18.4	19.2	20.1	20.8	21.6	22.4	23.2	23.9
360	+13.6	14.5	15.4	16.2	17.1	18.0	18.8	19.7	20.4	21.2	22.1	22.8	23.5
362	+13.2	14.1	15.0	15.9	16.8	17.6	18.4	19.3	20.1	20.9	21.7	22.4	23.1
364	+12.9	13.8	14.7	15.5	16.4	17.3	18.1	18.9	19.7	20.5	21.4	22.0	22.8
366	+12.6	13.4	14.3	15.2	16.1	17.0	17.8	18.6	19.4	20.2	21.0	21.7	22.5
368	+12.2	13.0	13.9	14.8	15.7	16.7	17.4	18.3	19.0	19.8	20.7	21.4	22.1
θ_e / W	14.0	14.5	15.0	15.5	16.0	16.5	17.0	17.5	18.0	18.5	19.0	19.5	20.0

Table 21

Meters per Second to Miles per Hour

(1 meter per second = 2.2369 miles per hour)

Meters per Second	0	1	2	3	4	5	6	7	8	9
	m.p.h.	m.p.h.	m.p.h.	m.p.h.	m.p.h.	m.p.h.	m.p.h.	m.p.h.	m.p.h.	m.p.h
0	0.0	2.2	4.5	6.7	8.9	11.2	13.4	15.7	17.9	20.1
10..........	22.4	24.6	26.8	29.1	31.3	33.6	35.8	38.0	40.3	42.5
20..........	44.7	47.0	49.2	51.5	53.7	55.9	58.2	60.4	62.6	64.9
30..........	67.1	69.3	71.6	73.8	76.1	78.3	80.5	82.8	85.0	87.2
40..........	89.5	91.7	94.0	96.2	98.4	100.7	102.9	105.1	107.4	109.6
50..........	111.8	114.1	116.3	118.6	120.8	123.0	125.3	127.5	129.7	132.0
60..........	134.2	136.4	138.8	141.0	143.2	145.4	147.7	149.9	152.1	154.3
70..........	156.6	158.8	161.1	163.3	165.5	167.7	170.0	172.2	174.4	176.7
80..........	178.9	181.2	183.4	185.7	187.9	190.1	192.3	194.6	196.8	199.1
90..........	201.3	203.6	205.8	208.0	210.2	212.5	214.7	0	219.2	221.5

Table 22

Beaufort Scale of Wind Velocities

Beaufort Number	Description of the Wind	Velocity	
		Statute Miles per Hour	Meters per Second
0......	Calm.........................	Less than 1.......	Less than 0.4
1......	Light air......................	1 to 3..........	0.4 to 1.5
2......	Light breeze....................	4 to 7..........	1.6 to 3.3
3......	Gentle breeze...................	8 to 12..........	3.4 to 5.4
4......	Moderate breeze................	13 to 18..........	5.5 to 7.9
5......	Fresh breeze....................	19 to 24..........	8.0 to 10.7
6......	Strong breeze...................	25 to 31..........	10.8 to 13.8
7......	Moderate gale..................	32 to 38..........	13.9 to 17.1
8......	Fresh gale......................	39 to 46..........	17.2 to 20.7
9......	Strong gale.....................	47 to 54..........	20.8 to 24.4
10......	Whole gale.....................	55 to 63..........	24.5 to 28.4
11......	Storm.........................	64 to 75..........	28.5 to 33.5
12......	Hurricane......................	Above 75.........	Above 33.5

Table 23

Density Correction Table for Upper Wind Charts

(This table may be employed to determine the approximate wind velocity from upper air pressure charts. Use Table 24 to determine the "gradient" wind, using the values of "*h*" and "*r*" obtained from the chart. Multiply this by the factors below to obtain the actual wind velocity at the elevation of the chart.)

Gradient level (600 meters)....................	1.00
5,000 feet...................................	1.10
10,000 feet...................................	1.28
15,000 feet...................................	1.50
20,000 feet...................................	1.77
25,000 feet...................................	2.10
30,000 feet...................................	2.52

Table 24

Gradient Wind Tables

[Giving velocity of gradient wind in miles per hour for given values of "*h*" (distance in miles between tenth inch isobars) and "*r*" (radius of curvature of isobars in km).]

A. Anticyclonic Circulation

h \ *r*	Latitude 30°													
	100	200	300	400	500	600	700	800	1000	1200	1500	2000	3000	∞
50	....	....	....	....	....	....	....	....	....	....	....	....	182	118
60	....	....	....	....	....	....	....	....	....	....	....	....	131	98
70	....	....	....	....	....	....	....	....	....	....	....	....	108	84
80	....	....	....	....	....	....	....	....	....	....	....	113	95	74
100	....	....	....	....	....	....	....	....	....	....	90	75	67	58
125	....	....	....	....	....	....	....	....	....	63	58	54	51	47
150	....	....	....	....	....	....	....	....	60	53	48	45	43	39
200	....	....	....	....	....	....	....	45	38	36	34	33	31	29
250	....	....	....	....	....	36	32	30	28	27	26	25	25	22
300	....	....	....	....	27	25	24	23	22	22	21	21	20	18
400	....	....	....	23	19	18	17	17	16	16	16	15	15	15
500	....	....	22	18	15	14	14	14	13	13	13	12	12	12
750	....	12	10	9	9	9	8	8	8	8	8	8	8	8
1000	....	9	7	7	7	6	6	6	6	6	6	6	6	6

h \ *r*	Latitude 45°													
	100	200	300	400	500	600	700	800	1000	1200	1500	2000	3000	∞
40	....	....	....	....	....	....	....	....	....	....	....	161	126	103
50	....	....	....	....	....	....	....	....	....	....	128	107	96	84
60	....	....	....	....	....	....	....	....	....	112	93	83	77	69
70	....	....	....	....	....	....	....	....	....	88	76	70	65	60
80	....	....	....	....	....	....	....	....	80	70	64	60	57	52
100	....	....	....	....	....	....	72	60	53	50	47	46	44	41
125	....	....	....	....	....	55	49	44	40	38	37	36	35	33
150	....	....	....	....	43	37	35	33	32	31	30	29	29	28
200	....	....	....	32	27	25	24	24	23	23	22	22	21	21
250	....	....	26	21	20	19	19	18	18	18	17	17	17	17
300	....	....	19	17	16	16	16	15	15	15	14	14	14	14
400	....	16	13	12	12	11	11	11	11	11	11	11	11	10
500	....	13	10	10	9	9	9	9	9	9	9	8	8	8
750	8	6	6	6	6	6	6	6	6	6	6	6	6	6
1000	6	5	5	4	4	4	4	4	4	4	4	4	4	4

h \ *r*	Latitude 60°													
	100	200	300	400	500	600	700	800	1000	1200	1500	2000	3000	∞
40	....	....	....	....	....	....	....	....	....	....	116	102	94	83
50	....	....	....	....	....	....	....	....	105	91	84	78	74	68
60	....	....	....	....	....	....	....	92	76	70	66	63	60	56
70	....	....	....	....	....	....	....	72	62	58	56	53	51	49
80	....	....	....	....	....	....	62	57	52	50	48	46	45	42
100	....	....	....	....	52	45	42	41	39	38	37	36	35	34
125	....	....	....	....	36	34	32	31	30	29	29	28	28	27
150	....	....	....	30	28	27	26	25	25	24	24	23	23	23
200	....	....	24	21	20	19	18	18	18	18	18	17	17	17
250	....	21	17	16	15	15	15	14	14	14	14	14	14	14
300	....	15	14	13	12	12	12	12	12	12	12	12	11	11
400	....	10	9	9	9	9	9	9	9	9	9	9	9	8
500	....	8	8	7	7	7	7	7	7	7	7	7	7	6
750	6	5	5	5	5	5	5	5	5	5	5	5	5	5
1000	4	4	4	4	3	3	3	3	3	3	3	3	3	3

TABLE 24—*Continued*

B. CYCLONIC CIRCULATION

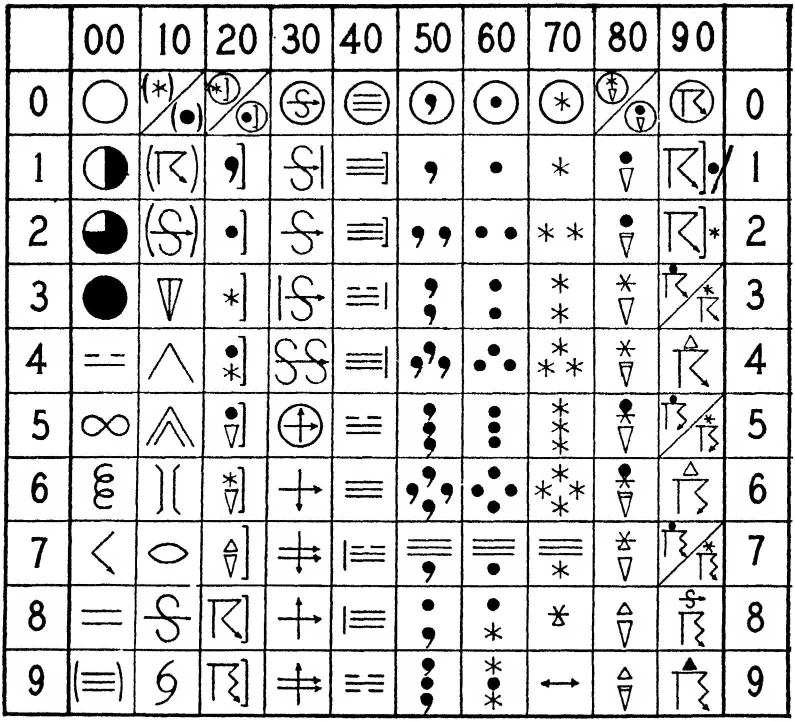
h \ r	Latitude 30°													
	100	200	300	400	500	600	700	800	1000	1200	1500	2000	3000	∞
30	54	71	83	92	99	105	110	115	122	128	134	144	154	193
40	41	54	63	70	75	79	83	86	92	96	102	108	116	145
50	37	49	56	62	66	70	73	76	80	84	88	93	99	118
60	33	43	50	54	58	61	64	66	70	72	75	79	84	98
70	30	38	44	48	51	54	57	60	61	64	66	69	73	84
80	27	35	40	44	47	49	51	52	55	57	59	62	65	74
100	24	31	35	38	40	42	43	44	46	47	49	51	53	58
125	21	26	30	32	34	35	36	37	38	39	41	43	44	47
150	19	23	26	28	29	30	31	32	33	34	35	35	36	39
200	15	19	21	22	23	24	24	25	25	26	26	27	28	29
250	13	15	17	18	18	19	19	20	20	20	21	21	21	22
300	11	13	14	15	15	16	16	16	17	17	17	17	17	18
400	9	11	12	12	13	13	13	13	14	14	14	14	15	15
500	7	9	9	10	10	10	10	11	11	11	11	11	12	12
750	6	7	7	7	7	7	7	7	7	8	8	8	8	8
1000	4	5	5	5	5	5	6	6	6	6	6	6	6	6

h \ r	Latitude 45°													
	100	200	300	400	500	600	700	800	1000	1200	1500	2000	3000	∞
30	51	65	75	81	87	91	94	97	102	106	110	115	121	137
40	38	49	56	61	65	68	71	73	77	79	82	86	90	103
50	35	44	50	54	57	59	61	63	66	68	70	73	76	84
60	30	37	43	47	49	51	53	54	56	58	59	61	63	69
70	27	34	38	41	43	45	46	47	49	50	52	53	55	60
80	25	31	34	37	39	40	41	42	44	45	46	46	48	52
100	22	27	29	31	33	34	34	35	36	37	37	38	39	41
125	19	23	25	26	27	28	28	29	29	29	30	31	31	33
150	17	20	21	22	23	24	24	25	25	25	26	26	27	28
200	13	15	17	17	18	18	19	19	19	19	20	20	20	21
250	11	13	14	14	15	15	15	15	16	16	16	16	16	17
300	9	11	12	12	12	13	13	13	13	13	13	13	14	14
400	8	9	9	9	10	10	10	10	10	10	10	10	10	10
500	6	7	7	8	8	8	8	8	8	8	8	8	8	8
750	5	5	5	5	5	5	5	5	5	5	5	5	5	6
1000	3	4	4	4	4	4	4	4	4	4	4	4	4	4

h \ r	Latitude 60°													
	100	200	300	400	500	600	700	800	1000	1200	1500	2000	3000	∞
30	48	61	69	74	78	81	84	86	90	92	95	98	102	110
40	36	46	52	56	59	61	63	65	67	69	71	74	77	83
50	35	41	45	48	50	53	54	55	57	59	60	62	64	68
60	29	35	39	42	44	45	46	47	48	49	50	52	53	56
70	26	32	35	38	39	40	41	42	42	43	44	46	47	49
80	23	28	31	33	34	35	36	36	37	38	39	40	40	42
100	20	24	26	27	28	29	30	30	31	31	32	32	33	34
125	17	20	22	23	24	24	25	25	25	25	26	26	26	27
150	15	17	19	19	20	20	21	21	21	21	22	22	22	23
200	12	14	14	15	15	15	16	16	16	16	16	16	17	17
250	10	11	12	12	12	13	13	13	13	13	13	13	13	14
300	8	9	9	9	9	9	10	10	10	10	10	10	10	11
400	7	7	8	8	8	8	8	8	8	8	8	8	8	8
500	5	5	5	5	5	5	6	6	6	6	6	6	6	6
750	4	4	4	4	4	4	4	4	4	4	4	4	4	5
1000	3	3	3	3	3	3	3	3	3	3	3	3	3	3

TABLE 25

INTERNATIONAL WEATHER SYMBOLS



ABBREVIATED DESCRIPTION OF SKY AND SPECIAL PHENOMENA

00 Cloudless (from no clouds up to but not including one tenth)
01 Partly cloudy (from exactly one tenth to exactly five tenths)
02 Cloudy (over five tenths up to and including exactly nine tenths)
03 Overcast (over nine tenths)
04 Low fog, whether on ground or at sea
05 Haze (but visibility greater than 1,000 m., 1,100 yards)
06 Dust devils seen
07 Distant lightning
08 Light fog (visibility between 1,000 and 2,000 m., 1,100 and 2,200 yards)
09 Fog at a distance, but not at station (or ship)
10 Precipitation within sight
11 Thunder, without precipitation at the station (or ship)
12 Dust storm within sight, but not at station (or ship)
13 Ugly, threatening sky
14 Squally weather
15 Heavy squalls } in last 3 hours
16 Waterspouts seen } in last 3 hours
17 Visibility reduced by smoke (industrial, grass or forest fires) or volcanic ashes
18 Dust storm (visibility greater than 1,000 m., 1,100 yards)
19 Signs of tropical storm (hurricane)

TABLE 25—*Continued*

PRECIPITATION IN LAST HOUR BUT NOT AT TIME OF OBSERVATION

20	Precipitation (rain, drizzle, hail, snow, or sleet)	
21	Drizzle	other than showers
22	Rain	other than showers
23	Snow	other than showers
24	Rain and snow mixed	other than showers
25	Rain shower(s)	
26	Snow shower(s)	
27	Hail, or rain and hail shower(s)	
28	Slight thunderstorm	
29	Heavy thunderstorm	

DUST STORMS AND STORMS OF DRIFTING SNOW

(Visibility less than 1,000 m., 1,100 yards)

30	Dust or sand storm	
31	Dust or sand storm has decreased	
32	Dust or sand storm, no appreciable change	
33	Dust or sand storm has increased	
34	Line of dust storms	
35	Storm of drifting snow	
36	Slight storm of drifting snow	generally low
37	Heavy storm of drifting snow	generally low
38	Slight storm of drifting snow	generally high
39	Heavy storm of drifting snow	generally high

FOG

(Visibility less than 1,000 m., 1,100 yards)

40	Fog	
41	Moderate fog in last hour	but not at time of observation
42	Thick fog in last hour	but not at time of observation
43	Fog, sky discernible	has become thinner during last hour
44	Fog, sky not discernible	has become thinner during last hour
45	Fog, sky discernible	no appreciable change during last hour
46	Fog, sky not discernible	no appreciable change during last hour
47	Fog, sky discernible	has begun or become thicker during last hour
48	Fog, sky not discernible	has begun or become thicker during last hour
49	Fog in patches	

DRIZZLE

(Precipitation consisting of numerous minute drops)

50	Drizzle	
51	Intermittent	slight drizzle
52	Continuous	slight drizzle
53	Intermittent	moderate drizzle
54	Continuous	moderate drizzle
55	Intermittent	thick drizzle
56	Continuous	thick drizzle
57	Drizzle and fog	
58	Slight or moderate	drizzle and rain
59	Thick	drizzle and rain

TABLE 25—*Continued*

RAIN

60	Rain	
61	Intermittent	slight rain
62	Continuous	slight rain
63	Intermittent	moderate rain
64	Continuous	moderate rain
65	Intermittent	heavy rain
66	Continuous	heavy rain
67	Rain and fog	
68	Slight or moderate	rain and snow, mixed
69	Heavy	rain and snow, mixed

SNOW

70	Snow (or snow and rain, mixed)	
71	Intermittent	slight snow in flakes
72	Continuous	slight snow in flakes
73	Intermittent	moderate snow in flakes
74	Continuous	moderate snow in flakes
75	Intermittent	heavy snow in flakes
76	Continuous	heavy snow in flakes
77	Snow and fog	
78	Grain of snow (frozen drizzle)	
79	Ice crystals, or frozen rain drops (sleet in the U. S. definition	

SHOWER(S)

80	Shower(s)	
81	Shower(s) of slight or moderate	rain
82	Shower(s) heavy	rain
83	Shower(s) of slight or moderate	snow
84	Shower(s) of heavy	snow
85	Shower(s) of slight or moderate	rain and snow
86	Shower(s) of heavy	rain and snow
87	Shower(s) of snow pellets (soft hail)	
88	Shower(s) of slight or moderate	hail, or rain and hail
89	Shower(s) of heavy	hail, or rain and hail

THUNDERSTORM AT TIME OF OBSERVATION

90	Thunderstorm	
91	Rain at time	thunderstorm during last hour, but no at time of observation
92	Snow, or rain and snow mixed, at time	thunderstorm during last hour, but no at time of observation
93	Thunderstorm, slight without hail, but with rain (or snow)	
94	Thunderstorm, slight with hail	
95	Thunderstorm, moderate without hail, but with rain (or snow)	
96	Thunderstorm, moderate with hail	
97	Thunderstorm, heavy without hail, but with rain (or snow)	
98	Thunderstorm combined with dust storm	
99	Thunderstorm, heavy with hail	

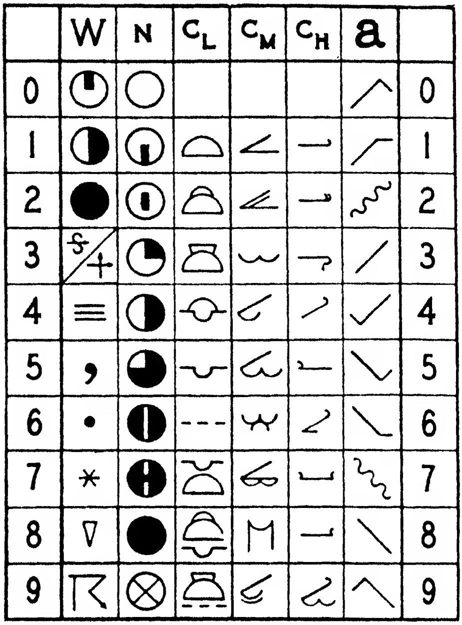


TABLE 25—*Continued*

W(PAST WEATHER)

0 Clear or a few scattered clouds
1 Partly cloudy
2 Cloudy to overcast
3 Sandstorm or duststorm, or storm of drifting snow
4 Fog
5 Drizzle
6 Rain
7 Snow or sleet
8 Shower
9 Thunderstorm

N(AMOUNT OF SKY COVERED BY ALL CLOUDS)

0 Absolutely no clouds in sky
1 Less than one tenth
2 One tenth
3 Two or three tenths
4 Four, five, or six tenths
5 Seven or eight tenths
6 Nine tenths
7 More than nine tenths but with openings
8 Sky completely covered with clouds
9 Sky obscured by fog, duststorm or other phenomenon

TABLE 25—*Continued*

C_L(LOWER CLOUDS)

0 No lower clouds
1 Cumulus of fine weather
2 Cumulus, heavy and swelling, without anvil top
3 Cumulonimbus
4 Stratocumulus formed by the flattening of cumulus clouds
5 Layer of stratus or stratocumulus
6 Low broken up clouds of bad weather
7 Cumulus of fine weather and stratocumulus
8 Heavy or swelling cumulus, or cumulonimbus, and stratocumulus
9 Heavy or swelling cumulus, or cumulonimbus, and low ragged clouds of bad weather

C_M(MIDDLE CLOUDS)

0 No middle clouds
1 Typical altostratus, thin
2 Typical altostratus, thick or nimbostratus
3 Altocumulus, or high stratocumulus; sheet at one level only
4 Altocumulus in small isolated patches; individual clouds often show signs of evaporation and are more or less lenticular in shape
5 Altocumulus arranged in more or less parallel bands, or an ordered layer advancing across the sky
6 Altocumulus formed by a spreading out of the tops of cumulus
7 Altocumulus associated with altostratus or altostratus with a partially alto-cumulus character
8 Altocumulus castellatus, or scattered cumuliform tufts
9 Altocumulus in several sheets at different levels, generally associated with thick fibrous veils of cloud and a chaotic appearance of the sky

C_H(HIGH CLOUDS)

0 No high clouds
1 Cirrus, delicate, not increasing, scattered and isolated masses
2 Cirrus, delicate, not increasing, abundant but not forming a continuous layer
3 Cirrus of anvil clouds, usually dense
4 Cirrus, increasing, generally in the form of hooks ending in a point or a small tuft
5 Cirrus (often in polar bands) or cirrostratus, advancing over the sky, but not more than 45° above the horizon
6 Cirrus (often in polar bands) or cirrostratus, advancing over the sky, and more than 45° above the horizon
7 Veil of cirrostratus covering the whole sky
8 Cirrostratus, not increasing, and not covering the whole sky
9 Cirrocumulus predominating, associated with a small quantity of cirrus

a(CHARACTERISTIC OF BAROMETRIC TENDENCY)

0	Rising, then falling	Barometer now higher than, or the same as 3 hours ago
1	Rising, then steady; or rising, then rising more slowly	
2	Unsteady	
3	Steady or rising	
4	Falling or steady, then rising; or rising, then rising more quickly	
5	Falling, then rising	
6	Falling, then steady; or falling, then falling more slowly	Barometer now lower than 3 hours ago
7	Unsteady	
8	Falling	
9	Steady or rising, then falling; or falling, then falling more quickly	

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